# ~~SOME ASPECTS OF~~ THE GLACIAL GEOMORPHOLOGY OF PARTS OF SOUTH WEST SHROPSHIRE AND ~~PART OF~~ MONTGOMERYSHIRE

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| | Thickness | Depth |
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| **CREW GREEN, BELLAN HOUSE (323157)** | | |
| Clay | 7-0 | 7-0 |
| Shale with gravel | 7-0 | 14-0 |
| Brown shale | 4-0 | 18-0 |
| Blue rock | 6-0 | 24-0 |
| **THE HILL, BAUSLEY, FORD** | | |
| Subsoil | 1-0 | 1-0 |
| Shale (blue) | 68-0 | 69-0 |
| Shale with marl layers | 51-0 | 120-0 |

## ABSTRACT

During the last glaciation the area was invaded by two distinct ice sheets. One, from the Irish Sea basin, pushed southwards across the Cheshire - Shropshire Plain and sent a lobe down the Vale of Stretton and onto the northern slopes of the Long Mynd. The other was nourished in the Plynlimon massif and advanced from the west.

Directions of ice movement, inferred from the orientation of ice-moulded landforms, drumlins, striations and the distribution of erratics, are complex, but the broad valleys were the main avenues of advance.

Drift generally occupies valleys and covers the lower hillslopes, while erratics and scattered patches of till are found at higher altitudes.

At the glacial maximum all save perhaps the highest hill summits were ice-covered. The Stiperstones tors are unreliable indicators of an unglaciated enclave.

Deglaciation was characterised by a gradual downwastage, retreat and stagnation of the margins of the ice. Complex systems of subglacial drainage channels were cut in the Vale of Stretton, and on all the major hill masses up to 1,300 ft.

Kame moraines deposited across the Onny valley at Lydham Heath and the Mule - Caebitra valley at Sarn, indicate pauses in the ice retreat. Kamiform topography is also a feature of the northern slopes of the Long Mynd and the Vale of Stretton.

Several important changes in the drainage pattern resulted from glaciation. The Mule and Camlad were diverted across pre-glacial watersheds through meltwater channels, while the Onny was diverted through an ice-breached col and meltwater channel at the southern end of the Long Mynd.

Radiocarbon dating of shells in the drifts at Iron Bridge, and of other drifts on the margins of the Welsh massif, indicate that the last glaciation of southern Shropshire was of Main Würm age.

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# Chapter I

# INTRODUCTION

## LOCATION AND SETTING OF THE AREA

The area under consideration comprises 674 $km^2$ of the central portion of the Welsh Borderlands (fig. I.I). Lying to the south-west of Shrewsbury, the irregular shaped area (fig. I.2) is bounded on the north-west by the Severn valley, on the south by the Kerry Hills and Clun Forest, and on the east by the Vale of Stretton. No physical feature defines the north-east boundary, but it generally includes that area above the 500-foot contour.

The region is dominated by several major hill masses, each separated by broad valleys. The hills, rising to over I,500 ft, include the Long Mynd, Stiperstones and Shelve district, the Long Mountain and Kerry Hills. Glaciation has resulted in a considerable derangement of the drainage and it is remarkable that in the centre of all the valleys except the Severn there is a low watershed from which the rivers flow in opposite directions. The principal rivers include the Severn, Camlad, Caebitra, Onny and Rea Brook.

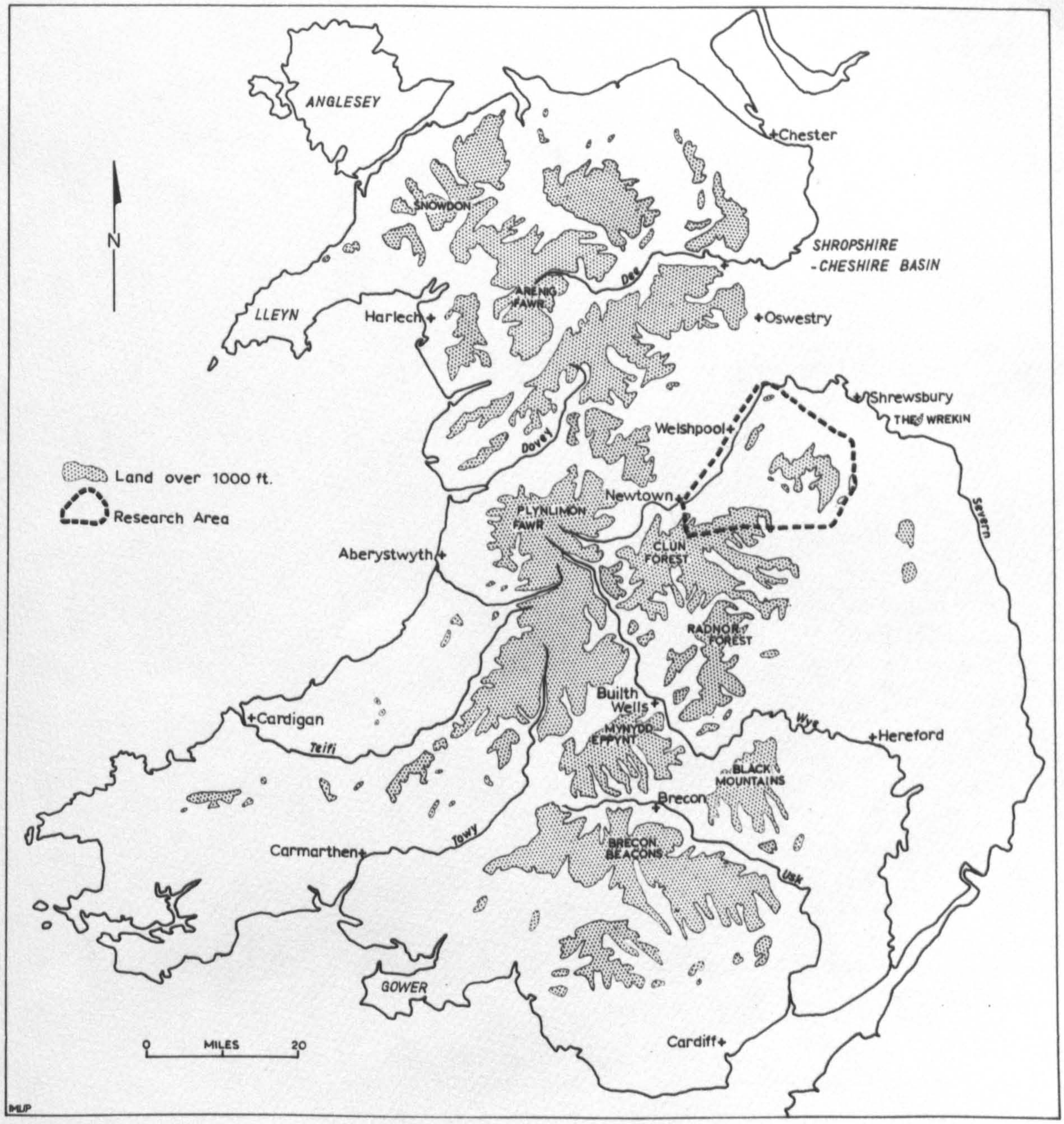


Fig. 1.1 Location of the research area

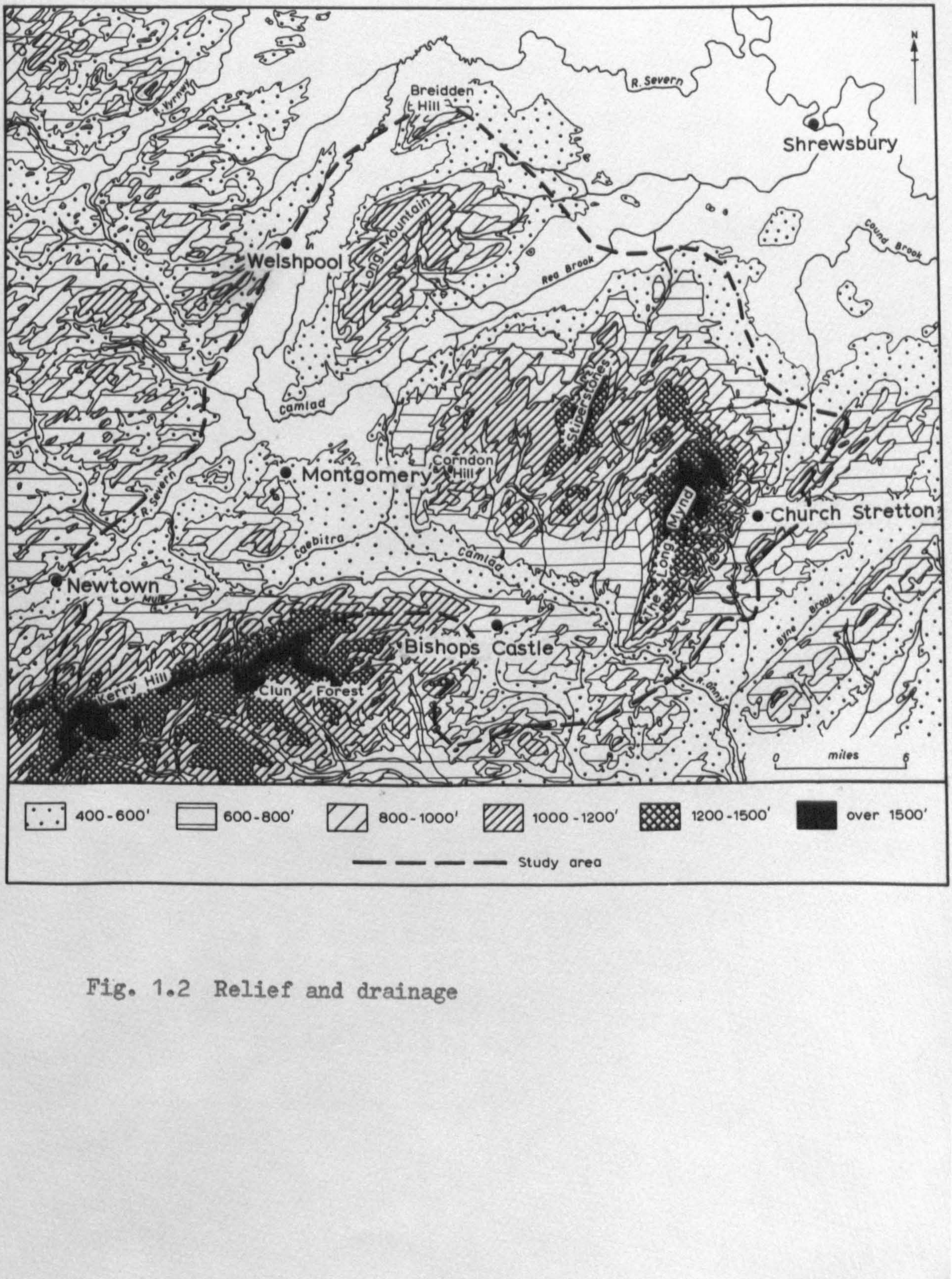


Fig. 1.2 Relief and drainage

# GEOLOGY

Detailed geological investigations have been made only in the Shropshire portion of the research area. Since the publication of Murchison's 'Silurian System'[1] this has become a classic area for the study of Pre-Cambrian and Paleozoic geology. Later workers include such renowned figures as Lapworth, Watts, Cobbold, Whittard, and the Geological Survey. Apart from a study by Earp of the Kerry Hills[2], the geology of much of the Montgomeryshire portion is known only from general surveys made by Murchisom and the Geological Survey.

Rocks of Pre-Cambrian to New Red Sandstone (Trias) age crop out (fig. I.3), though much of the area is dominated by the Pre-Cambrian and Paleozoic. The principal formations are listed in the Table below.

> "The district is remarkable for the great variety of its deposits for the immense range in time which these represent and for the number of unconformities...that occur at various horizons. These features bear testimony to the highly unstable conditions that have persisted throughout the geological history of the area..."[3]

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1. Murchison, R.I., 1839.
2. Earp, J., 1938.
3. Pocock, R.W., et al., 1938, p.5

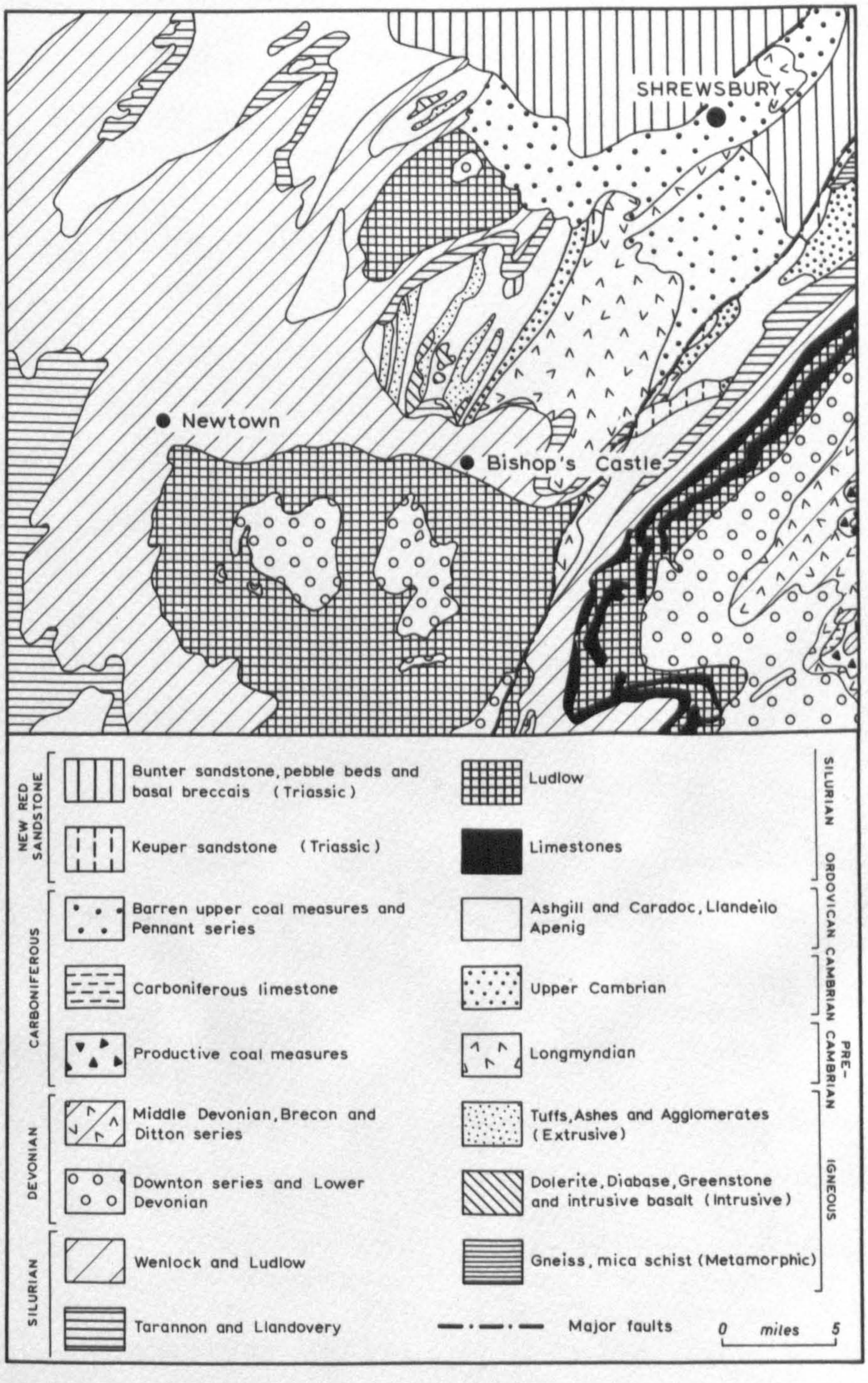



Fig. 1.3 Geology of the research area

## TABLE 1

### GEOLOGICAL FORMATIONS

| | | |
|---|---|---|
| RECENT and PLEISTOCENE | | Alluvium and Peat<br>Head deposits<br>Glacial till, sands and gravels. |
| TRIAS | Bunter | Upper Mottled Sandstone<br>Pebble Beds<br>Lower Mottled Sandstone |
| CARBONIFEROUS | Upper Coal Measures | Erbistock Group<br>Coed-y-allt Group<br>Ruabon Marl |
| DEVONIAN | Upper O.R.S.<br>Lower O.R.S.<br>Ditton Series<br>Downton Series | Farlow Series<br><br><br>Ledbury Group<br>Temeside Group |
| SILURIAN | Ludlow Series<br><br><br>Wenlock Series<br><br>Llandovery Series | Upper Ludlow shales<br>Aymestry Group<br>Lower Ludlow shales<br>Wenlock Limestone<br>Wenlock shales<br>Hughley (Purple) shales<br>Pentamerus Beds |
| ORDOVICIAN | Caradocian<br>Llandeillian<br>Llanvirnian<br>Arenigian | |

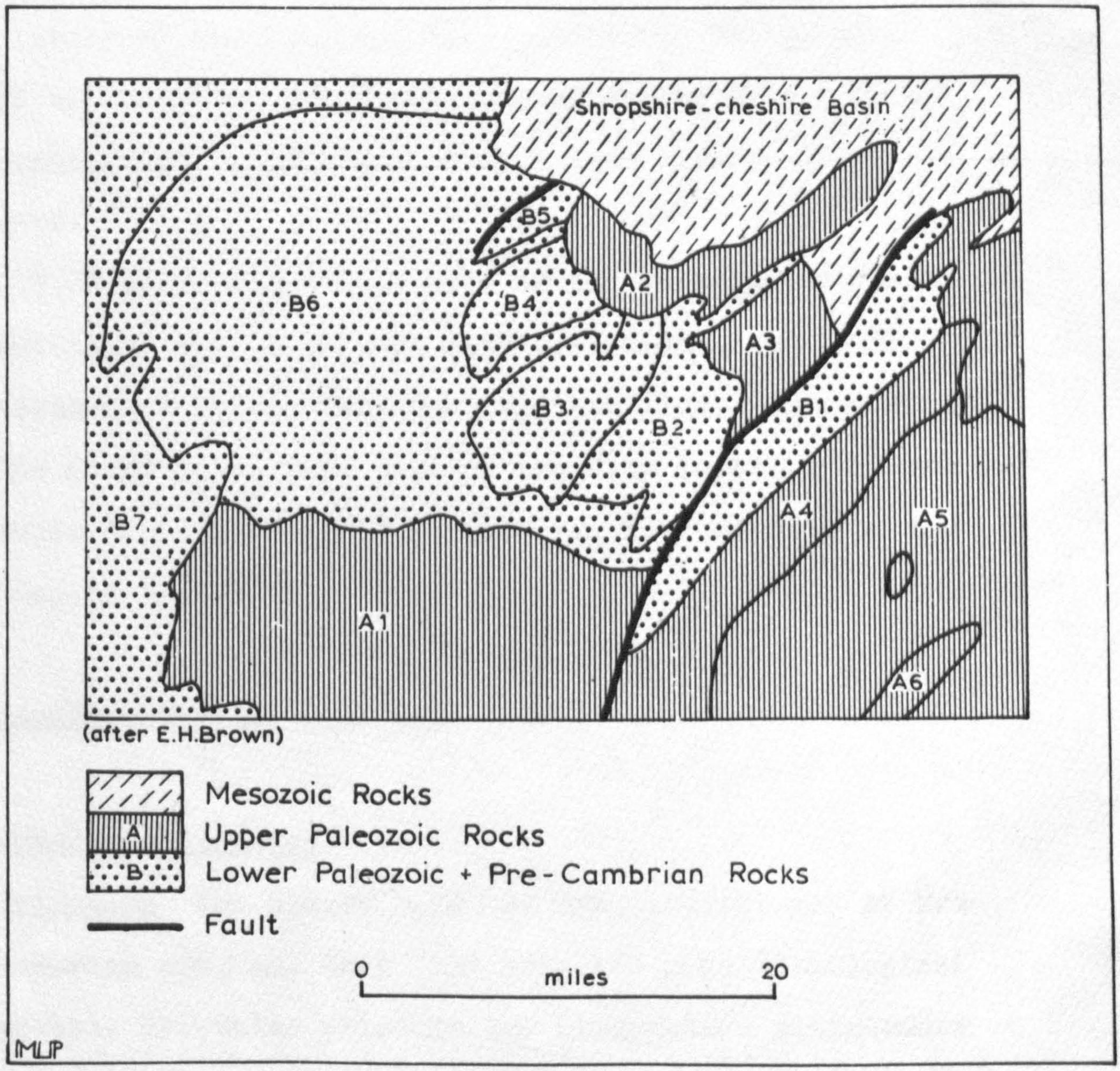
| | | |
|---|---|---|
| CAMBRIAN | Upper<br>Middle<br>Lower | Shineton shales<br>Upper Comley Series<br>Lower Comley Series<br>Wrekin Quartzite |
| PRE-CAMBRIAN | Longmyndian<br><br>Uriconian<br>Rushton Schists | Wentnor Series<br>Stretton Series |

The area has been unstable since the earliest times and has frequently found itself near a shoreline at the edge of deposition areas. Associated with this has been the continued recurrence of volcanic activity and folding and faulting on a considerable scale. The dominant trend of the rocks and of their folds and faults is from north-east to south-west, that characteristic of the Caledonian movement.

## Structure

The major structural elements of the region (fig. I.4) include the Cheshire-Shropshire Syncline, which influences the north-west portion, and the complex Long Mynd-Wrekin anticlinal uplift, which dominates the south-east. The principal divisions of the latter are the East Uriconian axis on the south-east and the Longmyndian ridge on the north-west, with the Leebotwood trough or basin between them. The East Uricomian axis comprises an extremely complex fault belt in which Uriconian and possible Longmyndian rocks are involved with Carboniferous and Triassic strata. The north-western boundary of the fault complex is formed by the Church Stretton Fault which runs south-west from the Vale of Stretton to Horderley and Aston-on-Clun. It appears to have had a marked effect on the geological

Fig. 1.4 Main Structural Elements




A1 - Clun Forest Basin
A2 - Hanwood Coalfield
A3 - Leebotwood Coalfield
A4 - Silurian Limestone Escarpment
A5 - Old Red Sandstone
A6 - Titterstone Clee Carboniferous outlier
B1 - Caer Caradoc - Wrekin uplift
B2 - Long Mynd Pre-Cambrian
B3 - Shelve anticline and Ritton Castle syncline
B4 - Long Mountain sysncline
B5 - Breidden hills anticline
B6 - Meifod Fold Belt
B7 - Central Wales Syncline

history of south Shropshire, and in the Church Stretton area it separates an extensive area of varied lithology and stratigraphy in the east from a less diverse region in the west.

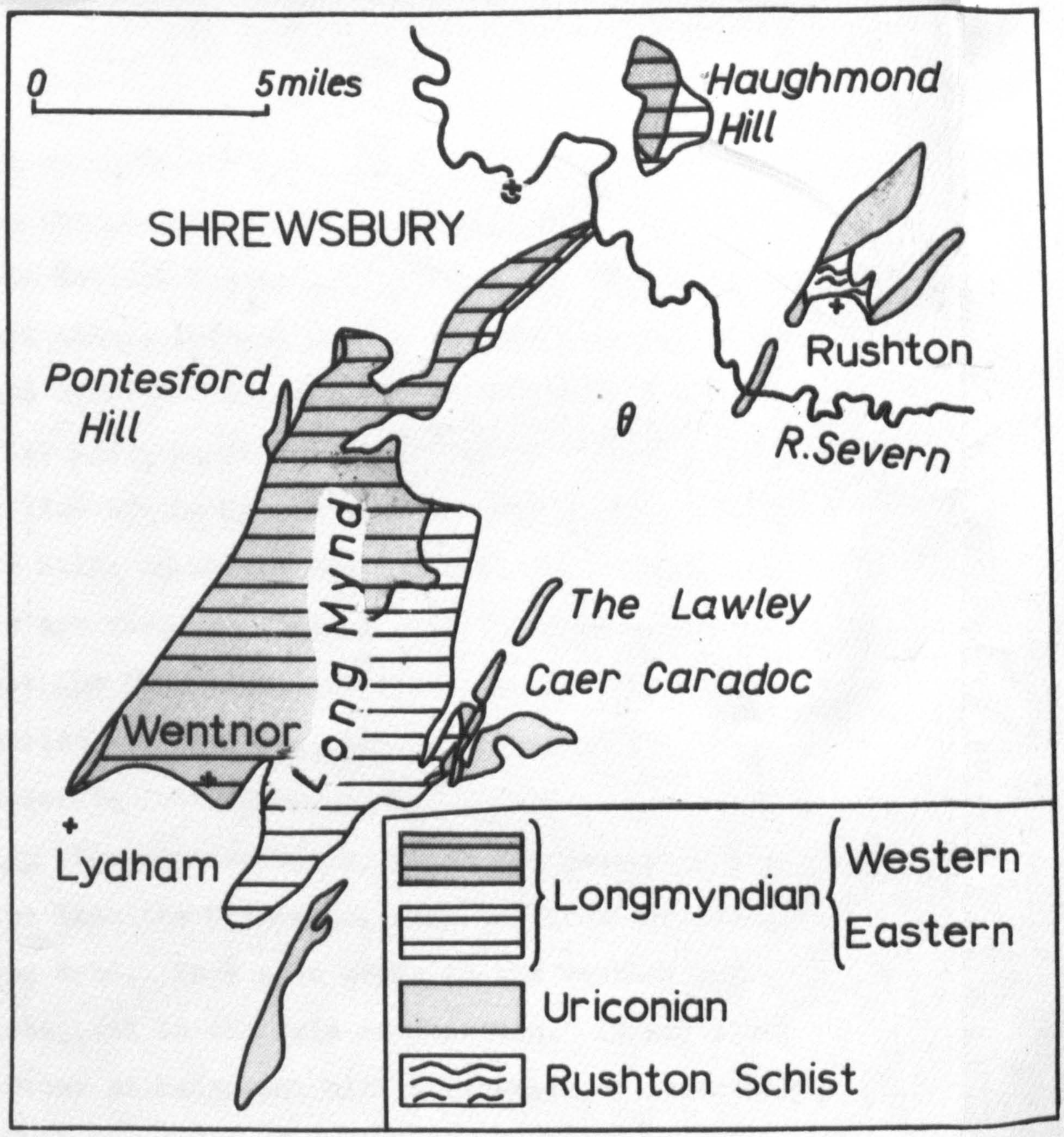
Bordering the Long Mynd on the west is the Shelve anticline and the Ritton Castle syncline, and to the north-west is the Long Mountain syncline. In the south, the Kerry Hills form part of the Clun Forest syncline, while the eastern half of the area is dominated by the complex Meifod fold belt.

## Stratigraphy and Lithology

### Pre-Cambrian Rocks

Uriconian The oldest rocks of the district are of Pre-Cambrian age, and they fall into two main lithological groups: Uriconian volcanic and Longmyndian sedimentary (fig. I.5).

The Uriconian rocks consist of lavas, volcanic breccias, agglomerates, and tuffs; dykes of felsite are common , and intrusive masses of pink granophyre invade the bedded series. Basic igneous rocks are found as dykes and sills penetrating the volcanic rocks. The outcrop of the Uriconian rocks occurs along two main lines




After Pocock et al

Fig. 1.5 Distribution of Pre-Cambrian Rocks in Shropshire

stretching north-east to south-west, one on the eastern and one on the western side of the Long Mynd: the Eastern and Western Uriconian respectively. The Uriconian massifs are always bounded on one or both sides by complex faults, and stand out as narrow, steep-sided hills. The Lawley, Caer Caradoc, Helmeth, Hazler, and Ragleth are examples along the line of the Church Stretton Fault, and Pontesford and Earl's Hills on the western side of the Long Mynd. The latter are composed chiefly of olivine basalts and dolerites. The East Uriconian hills are composed of a complex series of andesites, rhyolites and tuffs, with several dolerite intrusions.

Longmyndian Longmyndian rocks, which are generally considered to be older than the Uriconiam, form the prominent massif of the Long Mynd. They also occur in the Wentnor and Bridges area, and in the Vale of Stretton. In addition, outliers occur at Haughmond hill north-east of Shrewsbury, in the Cwms area of Caer Caradoc, in the Brokenstones, Horderley, and Wart hill areas, and on Hopesay hill.

The Longmyndian sequence comprises a series of diverse sedimentary rocks including grits, shales, siltstones and sandstones, varying from purple to greenish-grey in colour, with thin bands of tuff in some regions. The succession has been divided into two major parts:

the 'Stretton Series' (Eastern Longmyndian), which occupies the eastern part of the Long Mynd, and the 'Wentnor Series' (Western Longmyndian), which occupies the western part of the main outcrop. In general, the rocks of the Stretton Series are relatively fine-grained, while the Wentnor Series consist mainly of massive, often coarse, grits with beds of conglomerate, with subordinate shales and sandstone. The whole succession is probably one of shallow water deposition. The grits and conglomerates contain many fragments of volcanic rocks that closely resemble those of the Uriconian. Numerous dolerite dykes have been intruded into the sediments. They vary in thickness from a few feet to 200 ft, but are more commonly up to 10 ft across, and may be traced for up to half a mile.

According to James[1], the Longmyndian rocks are folded into a gigantic complex synclinal structure, with both limbs dipping steeply north-west. The core of the fold lies in the Bridges Group, roughly corresponding with the valley of the East Onny, and the axis of the fold trends north-north-east to south-south-west, parallel with the limbs.

The Rushton Schists are included in the Pre-Cambrian,

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1. James, J.H., 1956, p. 315-337

but do not crop out in the research area. They are found at Rushton, between the Wrekin and Chorlton Hill, and may have resulted from metamorphism of one on the other groups, or may belomg to a distinct and more ancient group.

## Cambrian

> "The deposition of the Cambrian sediments was preceded by the folding and prolonged denudation of the Pre-Cambrian rocks with the formation of an approximately level land surface over which a great transgression of the sea took place, in which the basal Cambrian deposits were laid down with strong unconformity on the older rocks." [1]

The rocks outcrop in two distinct areas. The major, eastern, area is associated with the East Uriconian axis, extending from Lilleshall and the Wrekin to the Caradoc and Comely. The minor, western, area is associated with the Western Uriconian axia and exhibits only the higher beds of the Cambrian, the Shineton (Habberley) Shales.

Lower Cambrian The Wrekin Quartzite flanks the eastern sdde of the Lawley, Caradoc and Cardington hills. The Lower Comeley Sandstone outcrop closely follows that of the Wrekin Quartzite, and is well-developed around Comeley, between Caradoc and the Lawley. This is a softer rock,

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1. Pocock, R.W. and Whitehead, T.H., 1948, p. 27

including sandstones and shales, and beds of glauconite.

Middle Cambrian The Upper Comeley Series consists of a series of conglomerates, breccias, sandstones and shales, outcropping in the Comeley area.

Upper Cambrian The soft, easily weathered shales of the Upper Cambrian form lower ground. They crop out in the gap between Comeley and Lawley, in the low ground east of Lawley and south-west of Shootrough. They are also found in the Habberley valley, between Habberley and Gatten, forming part of the eastern flank of the Stiperstones ridge.

Ordovician

The rocks of this system occur in four principal areas: the Stiperstones, Shelve and Chirbury area; the Breidden hills; the Pontesford area; and around Caradoc.

> "The Pre-Cambrian and Cambrian rocks seem to have formed...a land mass, against the western margin of which the Ordovician strata were deposited. The gradual sinking of this land allowed newer members of the system to be deposited successively further eastward..." [1]

Many of the rocks are fossiliferous. In the eastern areas of the outcrop they consist largely of shales, grits

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1. Op. cit., p. 38

and flags in which graptolites are common and are used to divide the strata into zones. In the Caradoc area the rocks are mainly arenaceous, with some shaley beds. At the Breiddens only Caradocian strata with volcanic rocks occur. The hills themselves are composed of volcanic rocks, equivalent to the Stapeley Volcanic Group, which are chiefly andesitic with a very distinctive green dolerite. The rock is dark greem, contains numerous pale-green feldspar crystals, and is readily recognised when found as erratics.

The fullest sequence is to be found in the Stiperstones, Shelve and Chirbury area, where the succession is nearly complete except for the absence of the Ashgill Series (fig. 1.6). The strata was divided by Lapworth[1] into a number of Groups (Stages), whilst the igneous rocks of the area have been made known largely by the studies of Watts[2]. The rocks have been thrown into wide folds, and the outcrops trend roughly from north-east to south-west. A syncline, more or less faulted, occurs at Disgwylfa (Squilver hill), and an anticline at Shelve hill and Corndon. The rocks are also cut by faults, which are often filled with intrusions of dolerite and picrite.

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1. Lapworth, C., 1879, p. 1-15; 1887. p. 78-80

2. Watts, W.W., 1885, p. 532-546; 1887, p. 670-671

The basic succession and lithologies of rocks in the Shelve district are shown on fig. 1.6; two are worthy of special mention. The Stiperstones Quartzite rests with an apparent conformity on the Shineton Shales. It is a much indurated quartzite and its sand grains are remarkably well rounded. It forms the central part of the Stiperstones ridge, where the quartzite stands out as tors and crags.

The Stapeley Volcanic Group (Ashes) represents a great volcanic outburst which may be correlated with the Llanvirnian activity. Away from the centre of the district shales subdivide the ashes into the Hyssington, Tasgar and Todleth Groups. They have been invaded by massive sills of andesites and dolerites. They form the prominent scarp of Stapeley hill and the masses of Llanfawr, Rountain and Todleth; in addition to this line they occur in a syncline to the east about Ritton Castle and Linley Park[1]. There is a remarkable coincidence between relief and geology. The volcanic rocks generally form the hills and ridges, and these are separated by broad valleys generally underlain by softer shales, all trending parallel to the strike.

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1. Watts, W.W., 1925, p. 342

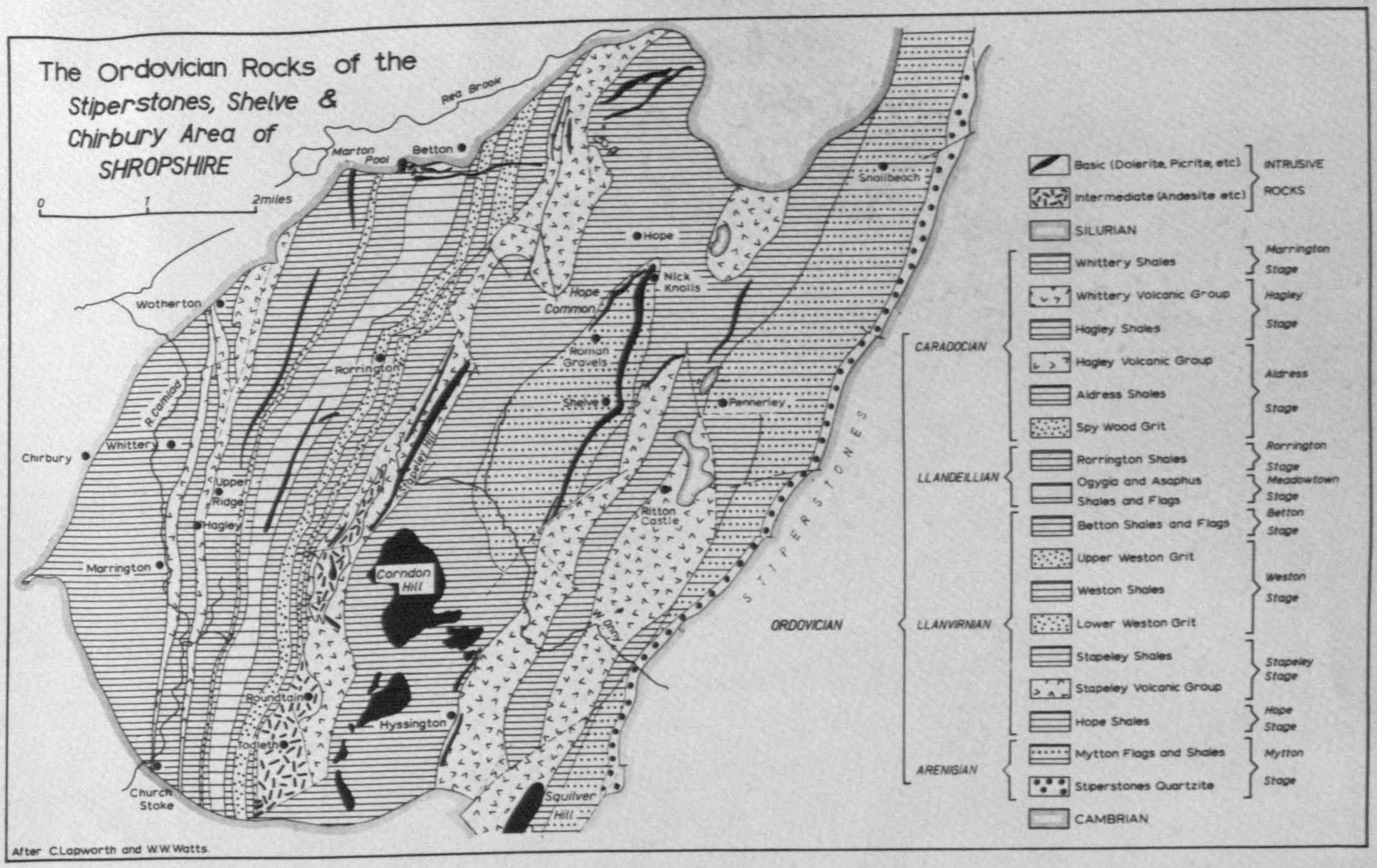


Fig. 1.6 Geology of the Shelve District

## Silurian

The sequence of rocks of the Silurian System was first established by Murchison[1] in the Welsh Borderland, a district that includes most of the localities after which the subdivisions were named. The rocks rest with strong unconformity on all divisions of the Ordovician, Cambrian and Pre-Cambrian, and individual beds show marked changes in lithology and fauna.

Silurian rocks occur in two chief areas of outcrop, separated by the Church Stretton Fault complex and the Eastern Uriconian axis. The rocks of the two areas (Western Outcrop and Main Outcrop) differ considerably in facies at some horizons.

> "The Shropshire type of Silurian rocks, mudstones alternating with limestones, is developed on the flanks of Wenlock Edge, stretching from the Severn at Bethnall Edge to Ludlow and beyond. But the Denbighshire type, flags and graptolitic mudstones, occurs also in the county, on the Long Mountain and Clun Forest."[2]

They also underlie much of the Montgomeryshire part of the district. There is a marked contrast between the shallow water reef limestone facies of the Wenlock Limestone and the graptolitic shales of equivalent horizon in the

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1. Murchison, R.I., 1839
2. Lapworth, C. and W.W. Watts, 1910.

Long Mountain area over 20 miles to the west.

The division into Valemtian, Salopian and Downtonian was first suggested by Lapworth[1], but the Silurian is now broadly divided into:

3. Ludlow Series
2. Wenlock Series
1. Llandovery Series

Llandovery Series IN Upper Llandovery times sandy or conglomeritic deposits were laid down along the irregular coastline of the partly submerged land-mass whose outstanding features were the Breidden, Shelve, Long Mynd, Caradoc, and Malvern hills. The most important area of land against which the deposits were laid down was the Long Mynd and Shelve districts. In the south-west of the Long Mynd they rest nearly flat against the steep western cliff of that plateau. Whittard has studied these deposits in detail[2] and they reveal such features as pebble beaches and sea stacks in the Upper Llandovery sea. The series has been subdivided into two groups:

Purple (Hughley) Shales
Pentamerus Beds

The Pentamerus Beds have a considerable variation in texture: conglomerated, pebbly grits, flinty sandstone, calcareous sandstone and occasional limestone to calcareous sandstone and mudstone. The outcrop closely follows the margin of the Long Mynd, from the Fault Scarp at

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1. Lapworth, C., 1879, p. 455
2. Whittard, W.F., 1932, p. 859-902

Plowden to Little Strettom and thence northwards to Church Stretton; they also occur along the margin of the Pre-Cambrian between Wentnor and Norbury, and adjacent to the Silurian outcrop on the eastern side of the Marton valley. In nearly all localities the rocks are characterised by fossils of Pentamerus oblongus, which are often abundant.

The Purple Shales occupy most of the low ground between Wentnor and the Long Mynd and southwards to Plowden. At the southern end of the Long Mynd the outcrop is displaced southwards by the Fault Scarp to the southern side of the Onny. From here the Shales have a continuous outcrop, largely obscured by drift, through Churchmoor Rough, Hamperley and Minton to the All Stretton area (fig. 1.7). They also occur on both sides of the Long Mountain, in the Middletown Gap, and the Marton valley.

Wenlock Series The Wenlock Series is best developed in the Wenlock Edge district, which lies outside the research area, though it has a limited development in the Western outcrop. It comprises two basic divisions:

Wenlock Limestone
Wenlock Shales

The Wenlock Shales consist mainly of flaggy and shaley grey siltstones and silty mudstones, locally

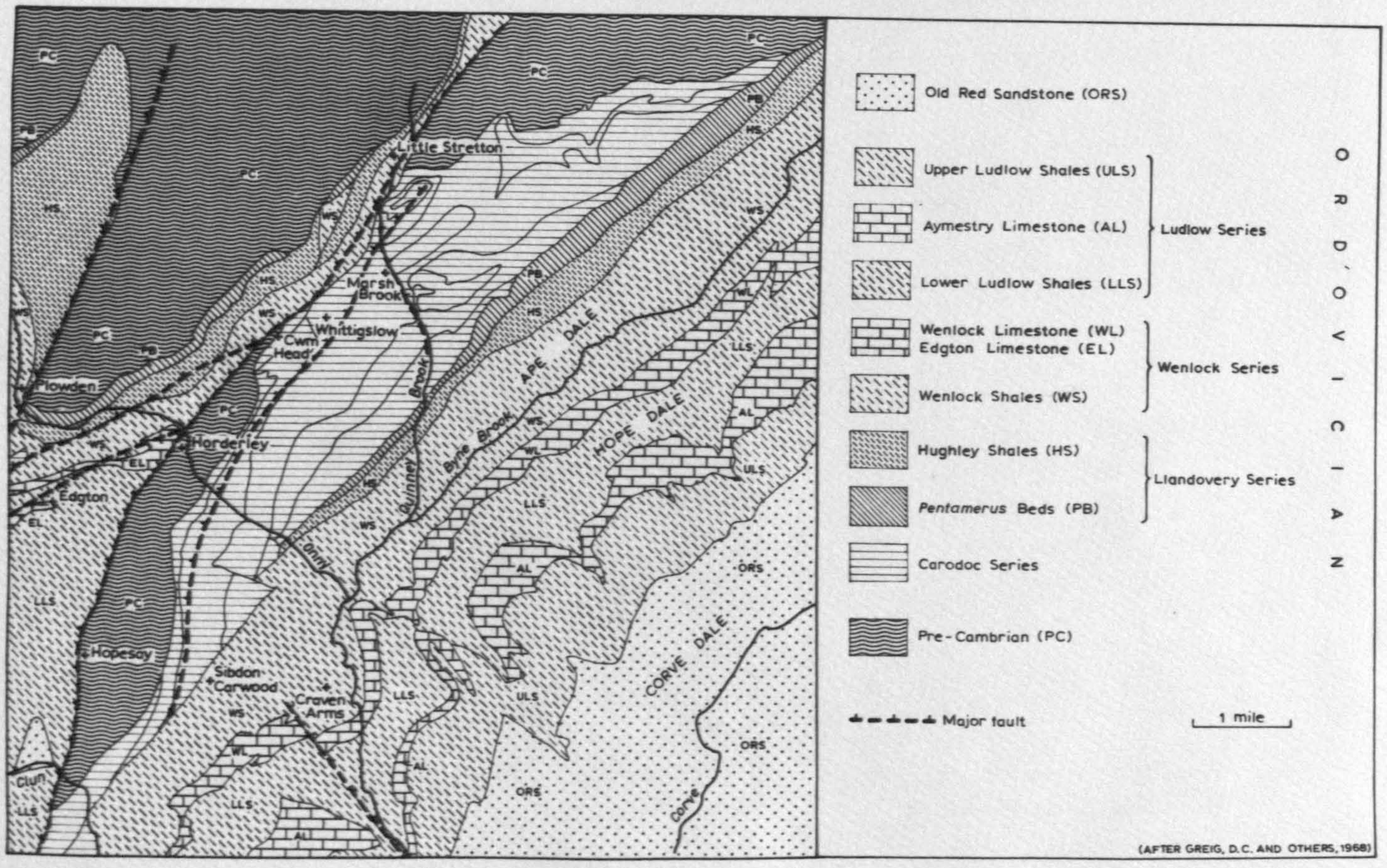


Fig. 1.7 Geology of the area between the Long Mynd and Corve Dale

calcareous, with thin nodular argillaceous limestone bands in the upper part and calcareous nodules towards the base. The Shales generally underlie the broad main valleys of the district, the Onny, Camlad, Caebitra and Rea Brook, though they are usually obscured by drift and alluvium.

There appears to be a transition from the Wenlock Shales to the Edgton Limestone, which extends from near Edgton towards Horderley, and caps the prominent scarp of Ridgeway hill. It consists of grey, thinly bedded, fine-grained argillaceous limestone, interbedded with bands of shaley mudstone.

Ludlow Series Rocks of the Ludlow Series have been broadly subdivided into three main groups:

Upper Ludlow Shales
Aymestry Group
Lower Ludlow Shales

They outcrop in the Long Mountain, Clun Forest and Church Stretton areas, but are best developed in the type locality at Ludlow. In the Long Mountain the Purple Shales of Middletown are succeeded by mudstomes, sandy mudstones and 'cardboard' shales, without any development of limestones. Earp[1] demonstrated the presence of Lower Ludlow shales, grits and flags in the Kerry Hills. In

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1. Earp,J., 1938, p. 125-160

the Church Stretton area the Ludlow Series is undivided and the outcrop extends south from the Edgton Limestone at Edgton and westwards from the Church Stretton Fault at Hopesay.

Old Red Sandstone (Devonian)

The main outcrop of the Old Red Sandstone is in the Corve Dale and Clee Hill districts. There is a smaller outcrop in the Clun Forest, and an outlier on the summit of the Long Mountain.

The classification adopted by the Geological Survey[1] is as follows:

| | |
|---|---|
| Upper Old Red Sandstone | Farlow Series<br>Clee Group |
| Lower Old Red Sandstone | Ditton Series<br>Downton Series |

The Downton Castle Sandstone and Temeside Shales of the Downton Series outcrop on the summit of the Long Mountain. The Devonian rocks of the Clun Forest area are restricted to the Downton and Ditton Series.

Carboniferous

Rocks of Carboniferous age occur in the north-east of the district, in the neighbourhood of Shrewsbury, and

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1. Greig, D.C. and others, 1968, p. 207

form the basis of the 'Shrewsbury Coalfields'. The pre-Carboniferous rocks of the Welsh Border district form the southern margin of the great synclinal basin of Shropshire and Cheshire, but it was not until Upper Coal Measure times that an extension of the basin took place, permitting deposition in the south. Thus, in the Shrewsbury Coalfields the Carboniferous Limestone, Millstone Grit, and the Lower and Middle Coal Measures are missing, and the Upper Coal Measures rest directly on Pre-Cambrian and Lower Paleozoic rocks.

The Hanwood Coalfield extends in a curved outcrop from Breidden hill to Haughmond hill. A little further east, the narrow Bayston hill ridge separates the Hanwood from the Leebotwood Coalfield. The little outlier of the Dryton Coalfield lies to the east of the main Church Stretton Fault.

Within the Upper Coal Measures, the Ruabon Marl outcrops only at the western end of the Hanwood coalfield, not further east than Westbury

Above the Ruabon Marl, the Coed-y-allt and Erbistock beds form the basis of the coalfields. It is within the former that the three workable seams of coal in the area occur.

Trias

Triassic rocks, forming part of the great basin of n north Shropshire and Cheshire, underlie the northern part of the research area. They are generally covered by drift, and are found in two areas separated by the Longmyndian ridge of Haughmond and Sharpstone hills. The only dividion of the Trias present is the Lower, or Bunter, comprising the Lower Mottled Sandstone, the Pebble Beds, and the Upper Mottled Sandstone.

The former, occupying most of the north-east area, is a soft non-pebbly sandstone, brick-red in colour, with greenish blotches and is usually highly false-bedded. The Pebble Beds consist of coarse sandstone, dull red or brown in colour, some pebbly and with lenses of conglomerate. The Upper Mottled Sandstone is nowhere exposed in the research area, but differs little from the Lower Mottled Sandstone.

## PRE-GLACIAL RELIEF

A study of the pre-glacial relief provides an essential backcloth from which to view the impact of the glacial epoch on the landscape; it also provides an important clue in the re-construction of the pre-glacial drainage, so that the extent of the derangement of the

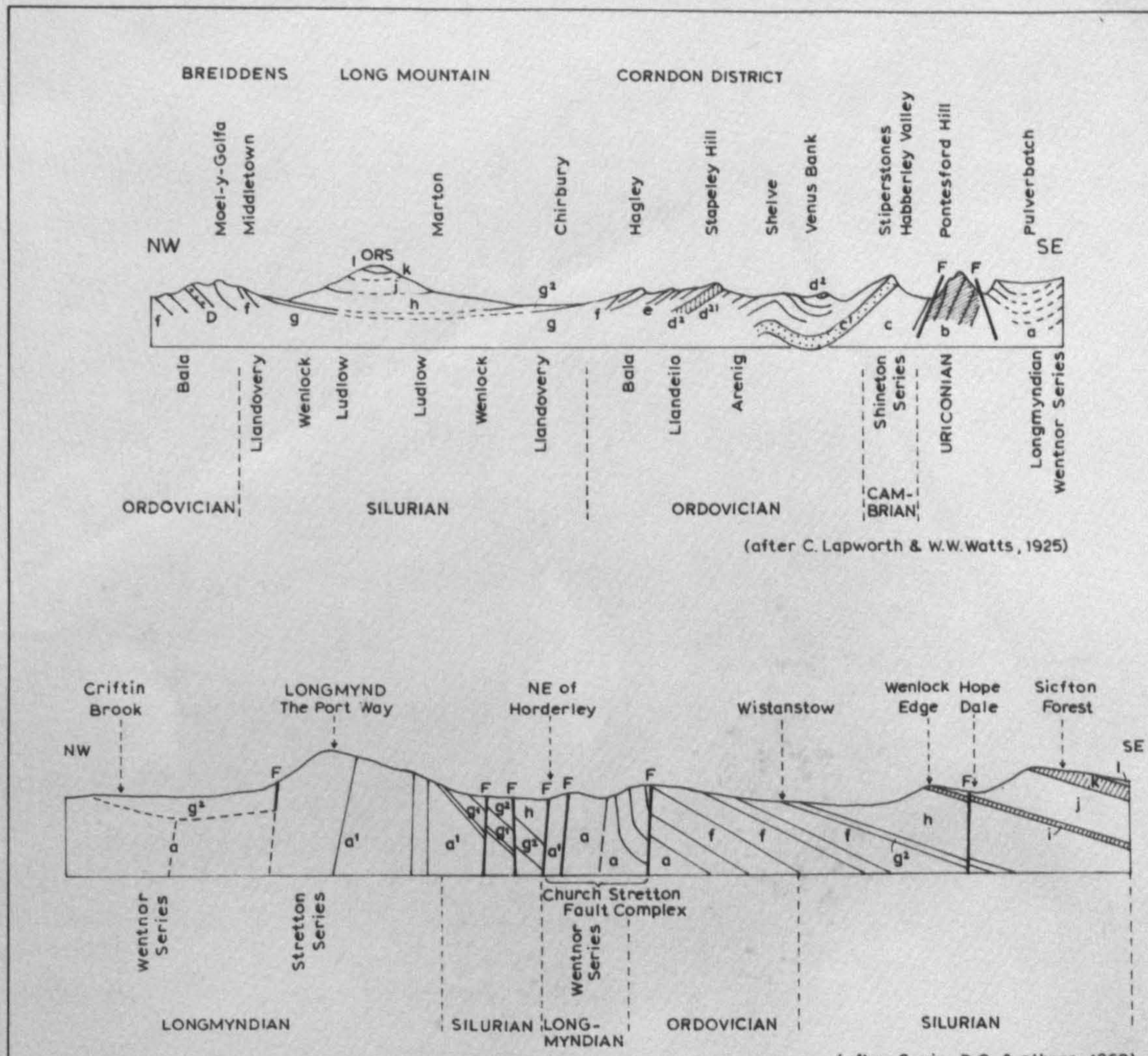


Fig. 1.8 Geological cross section between the Breiddens and Hope Dale

drainage during the glacial period can be determined. Topographically rounded and subdued landscape forms characterise the area; extensive plateaux bevel the summits of the principal mountains, and smaller plateau remnants are found at all elevations from 600 ft to 1,600 ft.

On the upper rolling hillslopes scareely a trace of glacial action can be seen; there is little evidence of large-scale erosion in the form of cwms, troughs or breached watersheds, and the deposits are generally thin and scattered. The mountain forms are substantially the result of pre-glacial denudation. It is not implied that they have been unvisited by ice, since there is evidence that most of the district must have been ice-covered, but that the ice passed over them and changed them but little. The effects of glaciation appear to be concentrated in the lowlands where deposits are often over 100 ft thick and where glacial erosion has modified the landscape to some degree.

Brown[1] demonstrated that a long series of plateau remnants found throughout Wales fell into three major height groupings: 700-1,100 ft, 1,200-1,600 ft, and

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1. Brown, E.H., 1960

1,700-1,900 ft. It was considered that these are representatives of former extensive planation surfaces, termed respectively the Low and Middle Peneplains and the High Plateau. Rising above the latter are a series of monadnocks, across the crests of which is an indication of a fourth high-level surface, the Summit Plain.

An examination of the plateau femnants found in this part of the Borderlands reveals that they fall generally into the major height groupings established by Brown.

Low Peneplain

Representatives of the Low Peneplain are found within all the major valleys and lowlands in the area. It is characterised by a high degree of dissection and a considerable irregularity of summit elevations, and it is possible to recognise two or more stages within the height range 700-1,100 ft.

In the Vale of Stretton the Long Mynd plateau is fringed by a group of flat-topped hills which progressively decrease in altitude towards the south, from 1,000 ft. at Church Stretton to 750 ft at Little Stretton. The latter may be traced along the edge of the Long Mynd towards Horderley. An extensive drift-covered surface at 650 ft exists above Marshbrook, at the southern end of the Vale of Stretton.

The Low Peneplain can also be traced on the south side of the Onny valley at the southern end of the Long Mynd, and along the broad Onny-Camlad valley. West of Montgomery, the heavily dissected hill mass has several summits which form part of the Low Peneplain. The most prominent is Goetre hill (1,000 ft) and Cefn-y-Coed (1,150 ft).

The Severn valley below Newtown is fringed by a strongly dissected landscape bearing traces of glacial erosion. Numerous ridges and hillocks show a rough accordance of elevation, ranging from 500-650 ft. The most extensive plateau remnant is at Forden, where a drift-covered surface is at 500-550 ft. Similar remnants can be traced down the Camlad-Rea Brook valley, along the lower slopes of Long Mountain.

Middle Peneplain

The Middle Peneplain is best preserved and most distinctive of the three erosion surfaces. It bevels the major upland areas, dominating the skyline, and is relatively little-dissected.

The gently undulating upland surface of the Kerry Hills forms part of the extensive residual planation surface cut across the Clun Forest basin. This rises westwards and northwards from Ludlow from 1,200-1,400 ft

to 1,600 ft at the Kerry Hills. Rising above the surface are a number of monadnocks, including Cilfaesty hill (1,732 ft).

The steep northern face of the Kerry Hills is dissected by a series of parallel valleys, on the interfluves of which are several plateau remnants. The Middle Peneplain is here divisible into at least two separate surfaces: 1,200-1,350 ft and 1,500-1,600 ft. The latter is well-seen on the summit of Kerry Hill, the former in the bench below, with a steep bluff between. The break between the two surfaces is also seen at the eastern end of the Kerry Hills, between Mason's Bank and Edenhope hill.

The steeply-folded Pre-Cambrian rocks of the Long Mynd are cut across by an extensive erosion surface between 1,350 ft and 1,500 ft (Pl. 1), divisible into at least two parts. A higher surface, at 1,450-1,500 dominates the central portion of the Long Mynd, above which Pole Bank stands as a monadnock at 1,695 ft. At the northern end of the Long Mynd is a lower surface at 1,300-1,350 ft, represented by Betchcott hills, Wilderley hill and Ratlinghope hill.

At both the northern and southern end of the Long Mynd the Middle Peneplain is truncated by a more steeply-sloping surface. In the south this may be part of the

Pl. 1 Middle Peneplain on the Long Mynd

Pl. 2 Middle Peneplain on Long Mountain

sub-Silurian surface, which has been described as a wave-cut surface with fossil stacks and beach deposits from exposures at the foot of the slope near the river Onny.[1] At the northern end, the proximity of the Triassic outcrop suggests the slope may be part of an exhumed sub-Triassic surface.

East of the Long Mynd, the summits of the chain of hills bordering the Vale of Stretton: Caer Caradoc, Lawley and Ragleth form part of the Middle Peneplain. To the west the flat-topped summits of several hills, including Linley (1,349 ft), the Gatten (1,300 ft), Heath Mynd (1,483 ft) and Black Rhadely hill (1,320 ft) carry the Middle Peneplain towards the Stiperstones. The northern and southern ends of the latter form part of the upper Middle Peneplain surface, whereas the central portion, at 1,731 ft, stands above it as a monadnock.

West of the Stiperstones the Middle Peneplain is clearly bevelled across a variety of rock types and a number of structures, including the Shelve anticline and the Ritton Castle syncline. Corndon hill, 1,684 ft, stands above the surface as a monadnock.

The gently undulating plateau surface of the Long Mountain falls northwards and eastwards from 1,338 ft to

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1. Whittard, W.F., 1932, p. 890

about 1,150 ft (Pl. 2), forming part of the lower division of the Middle Peneplain.

High Plateau

The High plateau is a highly dissected surface and is represented in this part of the Borderlands only as a series of monadnocks standing above the Middle Peneplain. The Kerry Hills surface is overlooked by the Cilfaesty hill monadnock. Further east the Stiperstones, Pole Bank and Corndon hill all rise above the Middle Peneplain. "These summits may collectively indicate a former extension of the High Plateau eastwards into England, without a notable gradient in that direction."[1]

## DRAINAGE

The present drainage system is dominated by broad subsequent valleys cut into relatively soft strata, generally Wenlock Shales, and by glacial meltwater channels in the floors of the pre-glacial valleys. The principal rivers include the Severn, Onny, Camlad, Caebitra Mule and Rea Brook. The pre-glacial drainage was probably also dominated by major streams running in the subsequent valleys, but these were articulated somewhat differently from the present system. The impact of glaciation has resulted in a considerable derangement of the drainage

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1. Brown, E.H., 1960, p. 91

so that deciphering the drainage evolution is an extremely complex task. Most of the major rivers now flow in opposite directions away from a low divide in the centre of the broad valleys. That between the Onny and Camlad is formed by a thick accumulation of glacial deposits at Lydham, as is that between the Camlad and Rea Brook at Marton.

Many of the rivers also cut across watersheds or across prominent hill masses. Always, this is the result of a permanent diversion of the drainage through a meltwater channel. The combined waters of the Camlad and Caebitra flow northwards through Marrington Dingle across a watershed, the river flowing against the rising hills. The Onny flows through a meltwater channel cut into the floor of the gap at the southern end of the Long Mynd between Plowden and Wistanstow. The considerable accumulation of drift at Sarn, and the cutting of a meltwater channel above Abermule, have resulted in the diversion of the Mule out of the broad Mule-Caebitra depression across the hill-mass west of Montgomery and into the Severn valley. There are many other peculiarities in the drainage system resulting from the glacial epoch, and these, together with those outlined above, will be discussed more fully in a later chapter.

# TECHNIQUES

Basic mapping of the area was done in the field using the 1:10,560 sheets of the Ordnance Survey. Particular note was taken of the colour, texture, erratic content and location of the glacial drifts, and drift boundaries were marked where possible. Erratics scattered above the limit of continuous drift are also significant and were recorded. A lack of quarrying within the area meant that fresh drift exposures were relatively few, and reliance often had to be placed on limited stream exposures. Augering was occasionally employed to determine the depth and character of the drifts, but this often proved too laborious for the results obtained and was not widely used. Much additional information on the depth and character of the drifts was gained by consulting numerous borehole records at the Institute of Geological Sciences, London.

Attention was also paid to the morphology of the drift in order to determine the character of glacial deposition and deglaciation. The numerous glacial meltwater channels throughout the area were mapped, and note was taken of their gradients and of their relationships to other glacial features.

Field work was supplemented by a study of aerial

photographs at a scale of about 1:10,000. In addition, a general survey was made of the glacial features of a wider area outside the research area.

In 1967, a survey of the glacial features of much of Wales was undertaken with my colleagues I.D. Ellis-Gruffydd and H.D. Foster, as a result of which much light was shed on the character, extent and age of the last glaciation. Radio-carbon dates were obtained from organic samples from within glacial deposits in numerous parts of Wales, including those at the entrance to the Iron Bridge gorge.

# Chapter 2

## GLACIAL EROSION

### INTRODUCTION

The importance of glacial erosion in modifying the landscape of the Welsh Borderlands has so far remained largely underestimated. Little reference to it is to be found in the published literature, where discussions of glaciation are largely dominated by considerations of glacial deposition and of deglaciation. This is perhaps because the influence of glacial erosion has not been so striking here as in other parts of Wales, such as Snowdonia, the Harlech Dome or the Brecon Beacons. These areas were essentially centres of ice dispersal[1] and cirques, U-shaped valleys, roches moutonnée and other erosional features are commonplace. Marked morphological contrasts exist between these areas and the Borderlands where much of the pre-glacial landscape remains intact. In South-west Shropshire remnants of at least three major erosion surfaces have been recognised and correlated with other parts of Wales, and the upland surfaces are rounded in outline. Rounded upland forms are generally regarded as remnants of a pre-glacial landscape, and not the result of glacial erosion.[2] The influence of glacial deposition, deglaciation and periglaciation in modifying the landscape of this area is readily apparent, but

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1. Foster, H.D., Ph.D Thesis, 1968. Ellis-Gruffydd, I.D., Ph.D. Thesis, in preparation.
2. Hollingworth, S.E., 1931, p. 286; Linton, D.L., 1969, p. 36-37.

there is some evidence also of glacial erosion. Rather than resulting in a wholesale modification, the effects have been more subtle and have added much detail to the landscape.

## CIRQUES

A cirque may be broadly defined as a

> "...deep, steep-sided recess, roughly semicircular in plan, cut into a slope by erosion beneath and around a bank of firn or a glacier."[1]

Cirques that are modified and enlarged by glaciers that head within them are true glacial cirques, as distinct from those excavated at the sites of banks of firn.[2] The latter have been termed nivation cirques by Russell.[3] These occur not only at the head of a valley, but also as a major indentation in an otherwise smooth slope on a valley or hill side.

Cirques may vary widely in form, depending on such factors as the pre-glacial form of the valley head, the regimen and duration of the glacier or firn bank, and, above all, the structure of the rock into which it was cut. However, it should be possible to distinguish glacial cirques from nivatioh cirques, or even cirques from unmodified valley heads, by certain diagnostic features.

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1. Flint, R.F., 1957, p. 97
2. Op.cit. p. 100
3. Russell, R.J., 1933, p. 936.

Nivation cirques are essentially shallow basins, produced by weathering, solifluction, and rill-wash beneath a bank of firn, the rock waste being moved downslope out of the cirque by mass-wasting. In some cirques coarse fragments slide down the firn bank and accumulate at its toe as a ridge, termed a protalus rampart.

> "Usually they are floored by periglacial deposits ..., display no overdeepening, and lack ridges of glacial origin."[1]

If accumulation of firn becomes sufficiently great flow will be induced and glacier erosion contributes to the further evolution of the cirque.

> "Rock fragments frost-wedged from the cirque walls are incorporated into the glacier and moved down its length, and an end moraine, differing from a protalus rampart in the wear of its constituent fragments, may be built at its terminus."[2]

The action of glacier ice abrades and quarries the cirque floor and may lead to the development of a rock basin of some considerable depth. After the disappearance of the ice these often contain small lakes.

However, although it is often relatively easy to distinguish the two types of cirque, in some cases this is difficult; there is no clear distinction between the change from firn to glacier cirques and the two forms may grade into one another.

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1. Peterson, J.A., 1968, p. 67
2. Flint, R.F., op. cit. p. 99-100.

Many of the upland areas in south-West Shropshire, such as the Longmynd, Stiperstones and Long Mountain, contain large abruptly-ending, amphitheatre-shaped valley heads which could be mistaken for glacial cirques. Rowlands claims that

> "The numerous large 'arm-chair' - like hollows along the west-facing slopes of Stapeley Hill, and to a lesser extent on Lan Fawr...",

support the possibility of a fairly widespread development of small glaciers in this region[1]. Similarly, Dwerryhouse and Miller claim that

> "... a small glacier flowed down the valley of the East Onny, at least as far as Plowden Station."[2]

Although no origin for this glacier is suggested, it would presumably have been nourished either in the Shelve area or on the west-facing slopes of the Longmynd.

The important question is whether these valley heads contain either true glacial cirques, or nivation cirques, and whether the area nourished any local ice.

As Rowlands states, several large hollows occur on the west-facing slopes of Stapeley Hill and the area between it and Lan Fawr to the south. The largest of these lies immediately west of Stapeley Hill (SO 308993). A close examination reveals, however, that this

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1. Rowlands, P.H., 1966, p. 49
2. Dwerryhouse, A.R. and A.A. Miller, 1929, p. 101

is not a simple 'arm-chair' - like hollow as Rowlands suggests. On the south-west side, the slope of the valley-head is relatively gentle and opens out onto Middleton Hill. It is dominated on the east by a prominent north-east trending ridge which rises abruptly above the floor. The feature shows no evidence of ice erosion and contains nothing resembling a moraine or protalus rampart. The sides and floor of this hollow are mantled with periglacial and post-glacial deposits; there is no trace of any glacial till.

To the south-east lies Middleton Beech (SO 303985) which has a somewhat 'hollow-like' form. Once again, however, the morphology of the feature is characterised by a series of north-east trending ridges across which it is incised. There is no sign of glacial erosion within the valley head; there is no moraine or till, and it is floored with a cover of head deposits.

A similar description could be applied to all the valley-heads and depressions along the west-facing slopes of this hill-mass, including those on Lan Fawr. These features do not have the appearance of glacial cirques, and it would be difficult to describe some of them as nivation features. Rather, they are to be interpreted as pre-glacial hollows; it seems possible that banks of firn would have occupied these sites at some stage of the glaciation and they may have undergone some modification by periglacial processes.

The west-facing slopes of the Stiperstones ridge are incised into by many deep and steep-sided valleys, the largest of which are Perkinsbeach, Myttonsbeach and Snailbeach. These all have abrupt, roughly semi-circular valley heads. Perkinsbeach

has two valley heads, each some 500 ft. high, incised into the Stiperstones plateau. Small gullies, probably of post-glacial origin, run directly down the valley sides and there are a number of bedrock outcrops. The valley floor is narrow but flat and is covered by head deposits, with some glacial drift at the lower end. Myttonsbeach (Pl.3) is of similar form, but has only one abrupt valley head, some 400 ft. high. There are many craggy rock outcrops on the valley sides and small gullies run down the headwall. Similarly, the valley floor is flat and oovered with head deposits.

A similar description could be applied to all the steep-sided valleys in this locality. Although the headwalls are roughly semi-circular and abrupt, none resembles a glacial cirque; there is no evidence of glacial erosion and there is no trace of moraine or protalus rampart. Nivational processes, including freeze-thaw and solifluction have probably contributed much to the development of these hollows, but there is no evidence that any nourished glacier ice during the Pleistocene.

Similarly, in the upland area to the east of the Stiperstones and on the western slopes of the Longmynd there are no features which resemble glacial cirques, and there are only a few forms which could possibly be termed nivation cirques. One of the latter is a semi-circular hollow some 250 ft. deep on the western side of Cothercott Hill (SO 408003); but again, there is no evidence of ice erosion and no till or moraine. Nivational processes were probably active during the Pleistocene, but the feature did not

Pl. 3 Myttonsbeach - amphitheatre valley head

nourish a glacier. Similar hollows occur further south on the west of Stitt Hill.

The eastern side of the Longmynd plateau is cut into by a number of steep-sided valleys, up to 500 ft. deep, locally termed 'batches'. The upper ends of some, for example Oakham Dingle, are large and combe-shaped, with an abrupt termination. These valley heads are rather similar in form to those described on the Stiperstones, with gullies running directly down them and with many craggy rock outcrops on the sides. Again, however, there is no evidence of glacial erosion; the valleys are narrow and winding with many craggy rock outcrops and abundant scree and head deposits. There is no sign of glacial till in the upper parts of these valleys.

The two remaining upland areas which could possibly have nourished glaciers, the Kerry Hills and the Long Mountain, similarly do not contain any forms resembling glacial cirques. The heads of many valleys in these uplands are 'amphitheatre'-shaped, but none bears evidence of ice erosion. Nivational processes may have contributed to their development, but none contain a protalus rampart.

In conclusion, therefore, it may be said that there is no evidence of any glacial cirques in the uplands of south-west Shropshire. The form of many arm-chair like hollows was probably influenced by nivational processes, but there is no basis for Rowland's claim of the fairly widespread development of small glaciers in this region. Also, as will be shown later, there is no basis for the claim by Dwerryhouse and Miller of a glacier moving down the East Onny valley, rather, there is evidence that the

ice moved northwards and not south in this region.

## ICE MOULDED LANDFORMS

In the lowland areas there is a widespread development of glacially-smoothed and ice-moulded landforms (fig. 2.1).

> "Many glaciated areas are characterised by groups of ridgelike and groovelike landforms... Though having a wide range of sizes and shapes, these forms apparently belong to a single family in that all are products of the streamline flow of glaciers, which moulded the subglacial floor through various combinations of erosion and deposition."[1]

The ridges and grooves may be composed entirely of glacial deposits or entirely of bedrock. The former are more generally known as drumlins, which are defined by Hollingworth as

> "... an elogate hill or mound of till possessing a smooth moulded outline. It is generally asymmetrical in long profile, and avoid in plan, being characterised by a steeper, blunter end, and a more gently sloping, tapering end."[2]

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1. Flint, R.F., 1957, p. 66.
2. Hollingworth, S.E., 1931, p. 325.

This is

> "... consistent with ice-moulding. The ice-ward end is generally broader, blunter, steeper and higher, and the distal end, narrower and more tapering... This asymmetry harmonizes with physical law ... they present their steeper face to the moving medium, in order to offer the minimum resistance to the flow by hindering the formation of vortices in the rear."[3]

Their long axes also parallel the direction of ice flow. Drumlins usually occur in groups, and in some places adjacent forms are indistinctly separated from each other. Compound drumlins occur in a variety of forms, including,

> "... double or triple ridges united at the steeper end with only the tails distinct; doublets en echelon, the tail of one rising from the flank of the other as an inclined terrace or shelf; small drumlins plastered on the sides of larger ones, giving a grooving effect to the flanks of the latter; two-tiered drumlins ..."[4]

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3. Charlesworth, J.K., 1958, p. 394-395
4. Hollingworth, S.E., op. cit., p. 326.

Text cut off in original

Not all drumlins are composed entirely of drift, and all variations occur from forms composed wholly of drift to thin-veneered solid types.

> " ... there is a complete gradation, independent of outward form and within a single field, from rock to drift."[1]

Those which consist almost entirely of bedrock are sometimes termed 'rock drumlins'.

In some cases large areas have been moulded by the ice into a series of rock drumlins. Thus Linton, referring to Selkirkshire, stated that:

> " ... bare ice-scoured ridges of Tarannon shales and greywackes emerge above drift-filled hollows. The former are commonly 100-300 yd. long and 30-50 yd. across, and are grouped in 'swarms' with their long axes trending between N 35° E and N 70° E; frequently they coalesce to form grooved and corrugated 'islands' of rocky ground surrounded by drift. They are usually only a few score feet high."[2]

Similarly, Hollingworth refers to the existence of considerable hills of rock in west Edenside that have been moulded into typical drumlin shapes that are parallel to, and frequently indistinguishable from the associated true drift drumlins.[3]

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1. Flint, R.F., op. cit., p. 68.
2. Linton, D.L., 1959, p. 39-40.
3. Hollingworth, S.E., op. cit., p. 341.

Ice-moulded landforms are common in the Severn valley between Newtown and the Breiddens, in the Mule-Caebitra valley, the Marton valley, and the Shelve area to the west of the Stiperstones.

Between Newtown and Abermule (SO 162946), on the southern side of the Severn valley, is a series of elongate bedrock ridges whose long axes trend roughly from southwest to northeast, parallel to the valley sides. They vary in size from those measuring 550 yd. long, 25-30 ft. high and 70 yd. across, to others 1000 yd. long, 15 ft. high and 30 ft. wide. They are often asymmetrical in long profile, being higher and steeper at the upvalley end. Good examples of these tapered ridges may be seen at Pen-y-gelli farm (SO 129917), Penarth Tack (SO 138916) and Llanmerewig (SO 159928). They generally have a thin covering of glacial drift, though bedrock outcrops are not infrequent. The ridges are separated by short, drift-filled, sub-parallel valleys which bear little relationship to the present drainage pattern.

Although by no means confined to lower levels, the ridges seem to be best developed below about 700 ft. At higher levels the tops of the larger hills seem also to have been modified by the passage of ice over them. Thus at Great Cloddiau (SO 159909), Weston (SO 143906) and Yewtree Cottage (SO 131907) are hilltops which are elongate in form and tapered to some degree.

The elongate hill ridges are continued northeastwards across the Mule valley towards Montgomery. Here again, they appear to be concentrated at lower levels, and are best seen west of Montgomery,

around Jamesford (SO 211971) and Sutton (SO 205968).

At Cefnbryntalch (SO 175963), on the southern side of the Severn valley, is a series of small bedrock ridges arranged en echelon, each being some 25 ft. high and 100 yd. long. About a mile downstream a number of prominent bedrock ridges occur in the centre of the Severn valley. The most prominent is that trending north-eastwards from Plas Meredydd (SO 185975). It is about three quarters of a mile long, 150 yd. wide and about 100 ft. high at the upvalley end, sloping down to about 25 ft. in the northeast. On either side of the ridge are other smaller ridges. West of Caer-Howel bridge (SO 197981) three of these, the surfaces of which are at about 325 ft., are cut across by the river Severn.

To the north of Montgomery, on the other side of the Camlad valley, similar elongated ridges occur around Forden (SO 228011). Few are visible on the broad plateau at about 500 ft., but they are particularly well-developed at about 300 ft. around Forden Station (SO 219006) and Brynhyfryd Hospital (SO 217001). Here a swarm of ridges occurs, each about 25 ft. high and 200 yd. long; all have a covering of glacial drift.

Although not studied in detail, elongate bedrock ridges are also a feature of the western side of the Severn valley. Examples of these are the ridges around Aber-Bechan (SO 142935), Dolforwyn (SO 162955), and Welshpool where they are particularly well-developed. The ridges are of similar form to those already described

and their long axes trend in the same general direction, parallel to the valley sides. All are covered with an unknown thickness of glacial drift, and most are confined to levels below 600 ft.

It seems probable that the formation of these ice-moulded landforms in the Severn valley has been strongly influenced by structure. Differential erosion by ice has emphasised the Caledonoid trend. Although structures in the Vale of Powis are not known in detail,

> " ... the Caledonoid trend of the valley is very striking and the folds and faults in neighbouring areas have a similar trend."[1]

In the Mule valley, to the southeast of Glan Mule (SO 162904), is a group rounded hillocks of drumlinoid form. They are 25-30 ft. high, 200-500 yd. long, are arranged with their long axes trending roughly from southwest to northeast and are higher and steeper at their southwest end. Although most of these features have a covering of glacial drift, it could not be ascertained whether they are composed entirely of drift or have a bedrock core. Numerous bedrock exposures on the surface of the ridge at Pen-y-bryn (SO 177901), indicate that this, at least, is an erosional form.

About one mile west-northwest of Sarn (SO 204908), on the hillslopes above the river Mule, is a group of isolated hillocks which occur right up to the watershed at Great Cefnyberin (SO 196919), lying between the Mule and the broad flat-floored valley which enters

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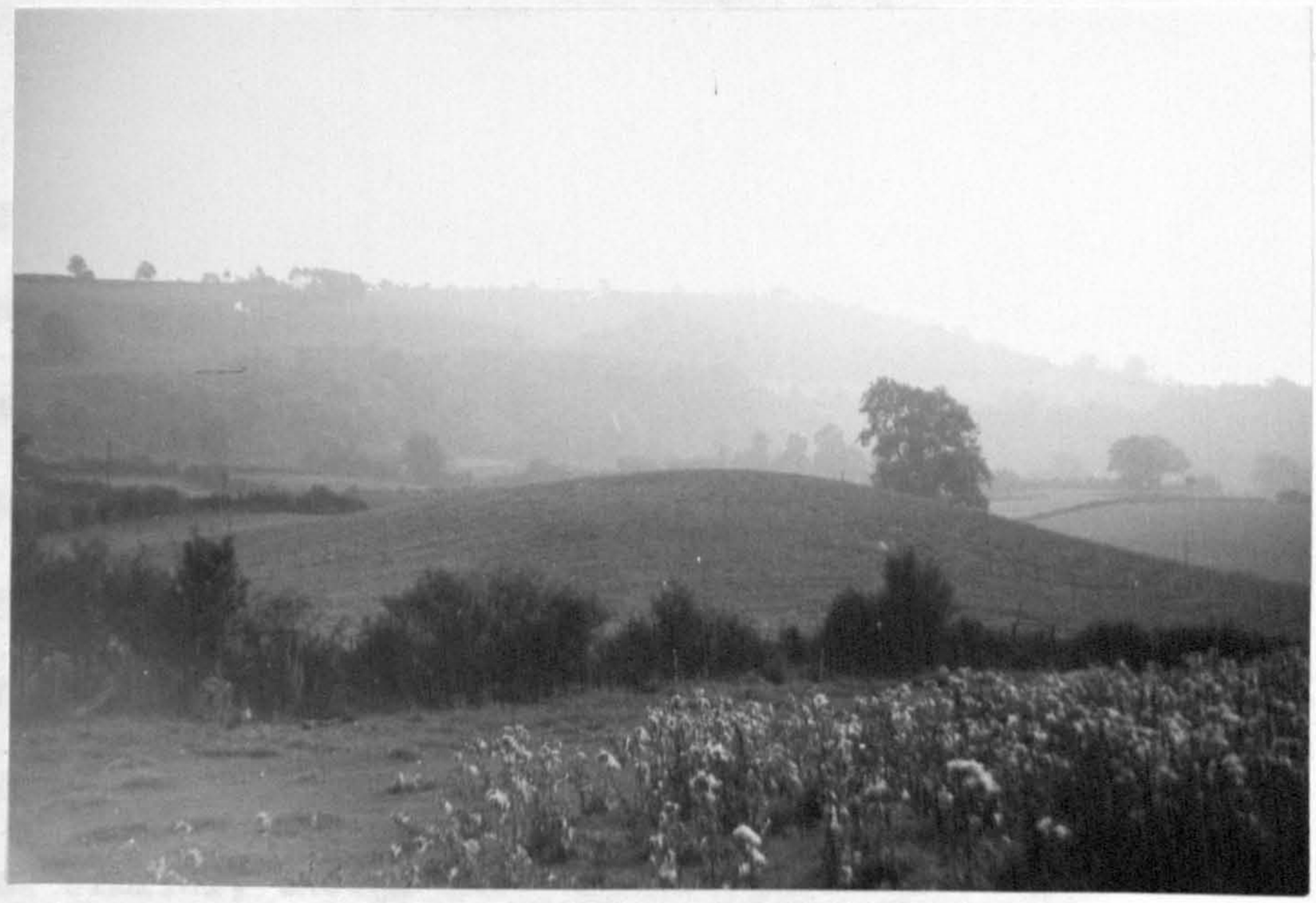
1. Brown, E.H., 1960, p. 131.

the Caebitra from the west. The hillocks are some 25-30 ft. high, 50-75 yd, long and their long axes are orientated roughly south-west-northeast. Most are covered with drift, but this is often thin and a quarry section in one to the north of Coed-y-brain (SO 191913) and numerous outcrops in others show them to be essentially bedrock features.

Immediately south of Sarn is an elongate ridge some 75 ft. high, which forms a continuation of the interfluve between the Mule valley and the small stream heading northeast from Shepherd's Lodge (SO 183884). The ridge is tapered in form, is roughly flat-topped at 700 ft., and although bedrock outcrops on the surface, drift is plastered quite thickly against the sides.

East of Sarn, on the southern side of the Caebitra valley, is another group of drumlin-shaped hillocks (Pl.4) rising out of the broad drift-covered plateau area north of Pentre Nant (SO 239912). The dimensions of these features vary, from the smallest measuring about 150 yd. long, 100 ft. wide and 30 ft, high, to others 700 yd. x 300 yd. x 60 ft. Although most are elongate, some are almost circular in plan. Their long axes are orientated southwest - northeast and they generally have a higher, steeper end facing upvalley. The drumlins are all covered with drift, but its thickness, or whether any have a bedrock core, could not be ascertained as only a few limited sections were observed.

In the upland area which lies between Chirbury (SO 262983) and Shelve (SO 337990), dominated by Corndon Hill, are numerous



Pl. 4 Drumlin in the Caebitra valley

Pl. 5 Drumlins near Pollardine, at the head of the Habberley valley

bedrock ridges trending roughly north-northwest to south-southeast., towards the Marton valley. Prominant among these is Ridge, which heads away from Hagley (SO 278976) for some two miles towards Rorrington Lodge (SJ 295006). This feature is some 600 yd. wide and 100 ft. high and although its lower portions are veneered with drift, it is essentially a bedrock form, being composed of Hagley volcanics. A number of isolated rock knolls occur on the crest and in some cases these resemble roches moutonnées, being smoothed and abraded on the southwest side but quarried and cliffed on the northeast; a good example is The Knoll (SO 282987).

A number of similar, though smaller, parallel rock ridges occur to the northeast of Priestweston (SO 291973), those on Stapeley Hill being a good example. They are 50-100 ft. high and are cut across, often at right angles, by transverse valleys, such as Cwm Dingle.

To the north similar bedrock ridges, trending in the same general direction, occur around Rorrington (SJ 301006). One, some 25 ft. high, heads away from the village for about three-quarters of a mile, while other, smaller features occur on the hillslopes to the east.

In this area, as in the Severn valley, the role of structure in the development of the ice-moulded landforms has been significant. The area to the west of Chirbury is underlain chiefly by Ordovician rocks (fig.16)which essentially comprises a series of Shales separated by beds of grits, flags, and volcanics. The latter, because of their greater resistance to erosion, usually form

upstanding hill-masses. A good example is Ridge which is composed of hard Hagley volcanics, and is bounded on either side by softer Hagley shales and Aldress shales. The dominant trend of the rocks and of their associated folds and faults, and hence of the upstanding ridges, is roughly from northeast to southwest. The ridges may well have formed prominent hill-masses before glaciation, but the ice in moving across the area has undoubtedly emphasised the features and probably also created some new forms.

On the northwest side of the Marton valley, between Marton (SJ 287015) and Westbury (SJ 357095), a broad, drift-covered undulating platform between 350-550 ft., rises above the flat valley floor and is backed by the steep edge of the Long Mountain. A number of hummocks and hills rise above the general level of the platform; they are generally elongate in form and their long axes trend roughly northeast - southwest, parallel to the valley side. The cover of drift is generally thin, and in many places bedrock was seen to occur, as for example on the ridge northeast of Binweston (SJ 302043). The ridges vary greatly in size, from the smallest, which generally occur in groups, measuring 600 ft. long, 150 ft. wide and 15-20 ft. high, to larger features, usually more isolated, 450 yd. x 600 ft. x 25-30 ft. Good examples of these ridges are to be seen northeast of Binweston, around Aston Pigott, Hawcocks and northeast of Causecastle Farm.

Marton stands on the northeastern end of a prominent ridge which rises abruptly to 400 ft. out of the flat valley floor.

The ridge is ¾ mile long, 400 yd. wide and 75 ft. high. It seems likely that this feature, like many of the smaller ridges, was largely fashioned by ice erosion.

On the southern side of the Marton valley, at Leigh Hall (SJ 334036) is a group of four small bedrock ridges of drumlinoid form and trending roughly northeast-southwest and forming a direct continuation of the ridges around Chirbury and Rorrington. At the northeastern end of the valley, around Minsterely (SJ 375051) and to the north of Ploxgreen (SJ 367037) is a series of hillocks and ridges, elongate in form, whose long axes trend parallel to the valley side. Most of these features are covered with glacial drift, and morphologically resemble drumlins in form; they occur in groups, and in one case a smaller drumlin has been formed on top of a larger one, giving a two-tiered effect. To the south of Reabrook (SJ 361045) two ridges occur as continuations of the hill-mass on the south of the Marton valley; bedrock outcrops indicate these to be erosional, rather than depositional forms.

An isolated occurence of drumlin-like features is to be seen on the eastern side of the Stiperstones some 700 yd. northwest of Gatten Lodge (SO 390992). A group of small knolls, evidently composed of glacial drift and with a northeast-southwest orientation, are arranged en echelon (Pl.6). They are each about 75 yd. long, 20 yd. wide and 15 ft. high and have an asymmetrical long profile, the steeper, blunter end pointing southwest.

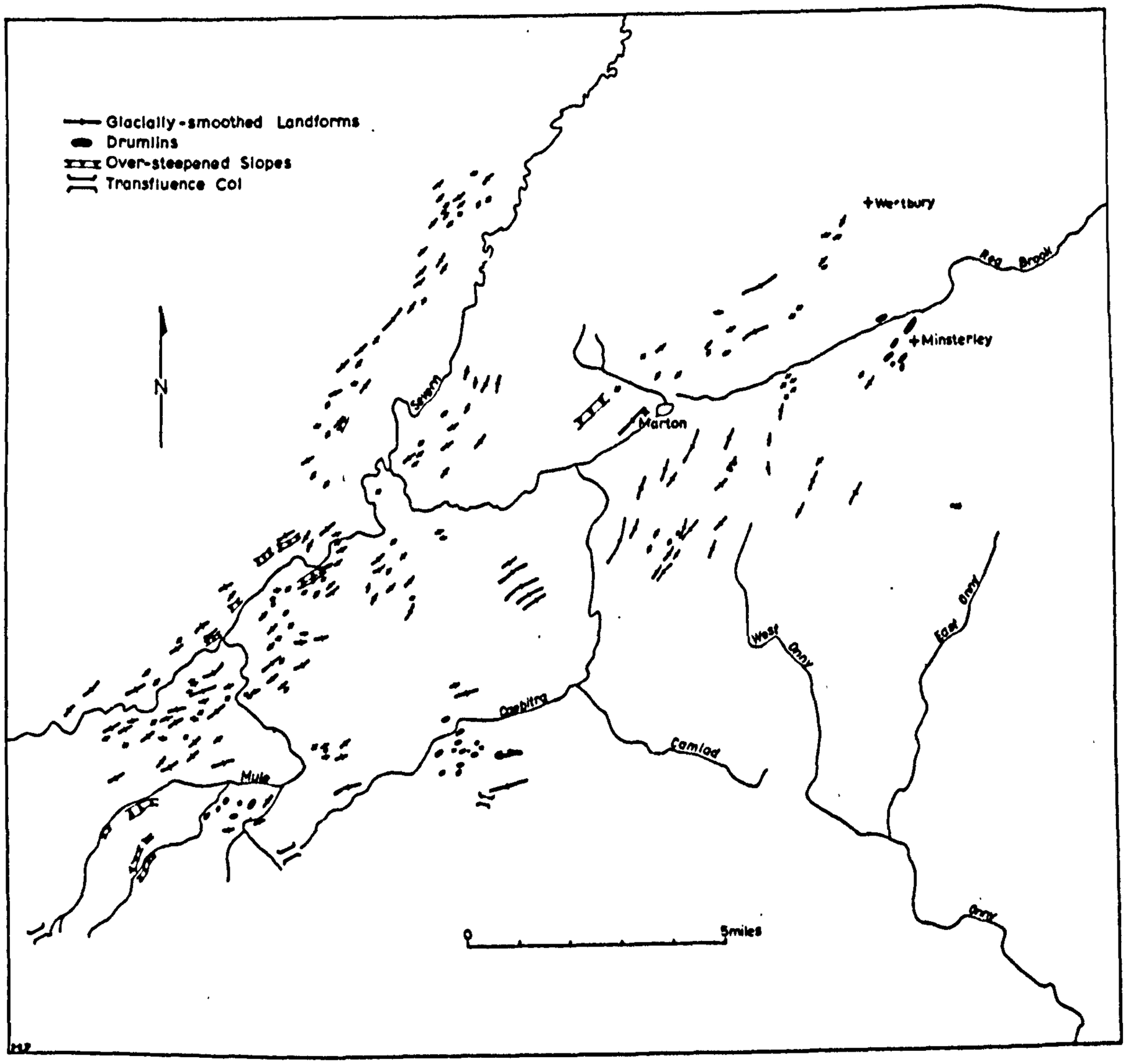



Fig. 2.1 Landforms of glacial erosion

The ice-moulded landforms occur principally at low elevations, on the margins of the largest valleys, and were probably formed beneath powerful streams of ice that moved down these valleys from the mountains of Central Wales. The intimate association of drift drumlins and rock drumlins is to be seen in both the Mule - Caebitra valley, south of Glan - Mule and east of Sarn, and the Marton valley, near Minsterley. Hollingworth considered that,

> "The existence of considerable hills of rock that have been moulded into typical drumlin shapes that are parallel to, and frequently indistinguishable from, the associated true drift drumlins, necessitates a prolonged and powerful ice movement in a constant direction."[1]

Evidence from meltwater channels, erratics and scattered till deposits indicates that ice reached up to at least 1,300 ft. on the Kerry Hills, Clun Forest and Long Mountain areas. Thus the ice in the Mule - Caebitra, Severn and Marton valleys must have been between 700 ft. and 1,000 ft. thick at the maximum of the glaciation. The ice-moulded landforms were probably formed beneath thick and swiftly moving ice which moved down the major valleys.

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1. Hollingworth, S.E., 1932, p. 341.

Evidence of ice-moulding on a much larger scale is to be seen on such prominent hills as the Breidden and Corndon, which are regarded as examples of crag-and-tail. It was proposed by Slater that the

> "... western or ice-ward flanks were directly exposed to the oncoming Welsh ice... The eastern or leeward flanks were zones of pluck..."[1]

Evidence for this is to be found in the contrasting morphology of the two sides of these hills and the distribution of glacial drift.

On the eastern, or leeward side of Corndon, are prominent crags and large areas of scree, often acres in extent, thickly strewn with massive blocks of dolerite which extend down onto the underlying Hope shales. By contrast on the western side such crags are conspicuously absent, the hill being rounded and smoothed in outline; and there is only a limited development of scree.

It is considered that the eastern flanks were zones of pluck, a process responsible for the disintegration and removal of the rock-face, aided by the downward inclination of the hard jointed dolerite. The western face was exposed to the abrading and smoothing action of the ice. In addition to scree, drift exposures reveal a 'tail' of drift on the eastern side of Corndon, extending across Hyssington Marsh,

Neighbouring hills, such as Roundton and Todleth, also show

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1. Slater G., 1944, p.164.

similarly contrasted topography on the two sides. The latter also exhibits, on its upper western and southern slopes a complex series of wide grooves, one of which to the north of the summit cuts across the crest of the hill. The grooves are up to 30 yd. wide and some 25-30 ft. deep. These were interpreted by Rowlands as glacial meltwater channels.[1] Similar, though less well-developed features occur on Roundton, and it seems likely that they are the product of ice erosion, rather than deglaciation.

## WATERSHED BREACHING BY ICE

Where glaciers occupy valleys from which there is no free outlet, the ice level in the valley will rise. This may occur either if the outlet is blocked by other ice, or if the level of the ice in the valley is rising more rapidly than the ice can escape to a lower level. The glacier surface may eventually rise above the lowest col in the watershed and the ice will escape across the col to a neighbouring area.

In the extreme west of the research area is a high watershed which extends roughly north-south and separates the drainage of the Mule and Teme from that of the Ithon. This is continued southwards by the Lugg-Arrow/Wye watershed. It is breached by a number of cols, all of which appear to be of pre-glacial origin and have been modified to varying degrees by the passage of ice through them.

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1. Rowlands p. 36-37.

The broad, flat col at Black Gate (SO 107853) is at about 1368 ft. and separates the source of the Ithon from that of the Mule. That ice passed through this col is demonstrated by the prominent accumulations of till at the head of the Mule. Within the col and on the hillslope to the north, are a number of elongate ridges each 25-30 ft. high, the long axes of which trend northeast to southwest, parallel to the sides of the col. They are veneered with drift, but occasional bedrock outcrops show them to be erosional forms. It seems likely that these are rock drumlins, fashioned by the ice as it passed over the watershed.

To the north, at Dolfor (SO 106871), is a larger col at about 1025 ft. Glacial drift within the col indicates that ice passed through it, but there is little evidence of glacial erosion.

The watershed between the Ithon drainage and the headwaters of the Teme is cut through by a small col at 1534 ft.[1] Again, the presence of till on the watershed indicates that it was crossed by ice, which evidently deepened and smoothed the sides of the col.

On the Kerry Hills, the interfluves separating the large northward-flowing streams are cut across by a number of cols, some of which show evidence of modification by ice erosion. For example, the broad, flat-floored col at Shepherd's Lodge (SO 184884), which is at a height of about 1050 ft. has glacial drift exposed in its floor and at the head of the valley to the northeast. The precipitous southern side of the col, which is some 125 ft. high

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1. Dwerryhouse, A.R. & A.A. Miller, 1930, p. 103.

and quite smooth and regular, is probably the product of ice erosion.

Further east, the broad col at Argoed Lodge (SO 244904) is floored by accumulations of glacial drift and there are prominent meltwater channels on either side. Again, it appears that ice passed through this col and that it has undergone modification in the form of a steepening of the sides.

In addition to a modification of pre-glacial cols, it seems probable that many of the valleys have been overdeepened and their sides steepened by the passage of ice through them. The southern side of the Mule valley around Gilfach (SO 135896) has a regular steep slope some 300 ft. high. Similarly, to the east, both sides of the Miheli valley, downstream from Middle Cwm-y-ddalfa (SO 138876) appear to have been oversteepened by ice.

Borehole records from the centre of the Mule valley at Kerry (SO 143901) and further downstream at Pen-y-gelli (SO 176905) reveal glacial clays and sands to a thickness of over 100 ft. and 104 ft. respectively.[1] In no case was the bedrock floor reached, and it seems possible that the considerable depth of the valley at these points is due to overdeepening by ice. Further downvalley, the steep regular hillslopes at the Highlands (SO 214904), and to the south of Sarn at Gyfennt Wood (SO 182895), may also be the product of ice erosion.

Since the Severn valley is bordered on either side by hillslopes that have evidently been smoothed into a series of rock

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1. See Appendix 1.

drumlins, it is to be expected that the valley itself should contain evidence of glacial erosion. Oversteepened hillslopes occur at several places on either side of the valley, for example at Abermule (SO 162945), at Fron to the West of the prominent elongate ridge in the centre of the valley, to the northeast of Berriew (SO 188008), and at the Moors (SJ 244093) to the northeast of Welshpool. It is also highly likely that the precipitous, 1000-ft. high northwest face of the Breidden and the southwest side of Moel-y-Golfa have been fashioned to some extent by ice erosion. There is little evidence of overdeepening of the Severn valley by ice, but the recording of severely shattered shales some 6-7 ft. beneath clay at Abermule[1] is perhaps evidence of some glacial erosion on the valley floor.

In the Marton valley, apart from the glacially-smoothed hillslopes, there appears to be little evidence of glacial erosion. However, the steep, regular, upper slopes of the valley to the west of Marton, and the southwest side of Aston Hill (SJ 324068), may have been fashioned by ice.

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1. See Appendix 1.

# CONCLUSION

The forms of erosion in peripheral areas of glaciation are not so dramatic as those at the centres of ice disposal. The main effect appears to be the creation of ice-moulded landforms, concentrated in the valleys, and, at higher levels, the breaching of watersheds by ice. There is little evidence of glacial erosion in the interfluve areas and on hill summits; the ice sheet appears to have been a more effective agent of erosion in the lowlands. This may be because the high-level areas are more resistant to erosion and because ice thicknesses were greater in the lowlands. The ice-flow was concentrated in the valleys, in such a way that the ice on the hill-tops was not able to move effectively.

# Chapter 3.

## DISTRIBUTION OF THE GLACIAL DRIFTS

### INTRODUCTION

The complex deposits of glaciers and ice-sheets are known under the general term 'glacial drift',

> "... which embraces all rock material carried in transport by glacier ice, all deposits made by glacier ice, and all deposits predominantly of glacial origin made...in bodies of glacial meltwater..."[1]

In the past, drift was divided into two supposedly distinct types: unstratified drift, or till (also sometimes known as boulder clay), and stratified drift. The former is a glacial deposit in which the selective activity of water played a minimal part in deposition. By contrast, stratified drift is often distinctly sorted according to size and weight of its fragments, indicating the presence of water during deposition.

In recent years, however, it has become increasingly clear that there is no distinct division between till and stratified drift, and that the one grades into the other. The more living glaciers are studied, the more easily it is to see that only gradation exists.

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1. Flint, R.F., 1957, p. 108.

> "The till is merely the end member of the stratified drift series and grades into this; for all till was deposited from the ice in the presence of at least a thin film of water."[1]

Also, water often washed out the finer particles from the drift during transport, and so left a sandy till.

> "Sands and gravels, intercalated with boulder clay, form nests, pockets or impersistant seams, sometimes distorted or ripple-marked."[2]

Because of this gradation till is now regarded as drift which is dominantly nonsorted according to grain size.

> "It is the non-size-sorted end member of a series whose opposite member is well-size-sorted stratified drift."[3]

The two most obvious characteristics of glacial till are absence of stratification and lack of size-sorting; it is physically and lithologically heterogeneous. Also, the texture of the till varies greatly. Thus Geikie stated that,

> "The deposit known as till or boulder clay, is usually a firm, tough tenaceous, stony clay, which gives every evidence of having been subjected to great pressure."[4]

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1. Charlesworth, J.K., 1959, p. 378.
2. Ibid.
3. Flint, R.F. Op. cit.
4. Geikie, J., 1877, p.7.

In the other regions, however, it often becomes coarser and sandier;

> " ... in certain districts it might be described as coarse agglomerate of subangular and angular stones set in a scanty matrix of coarse earthy grit and sand..."[1]

The character of the till is largely determined by the nature of the bedrock or superficial material over which the ice has passed, since both boulders and matrix are predominantly gathered from the underlying rocks.

> "Generally, the bulk, say 75-90% of the boulder clay, has not journeyed more than 50 miles,"[2]

and Krumbein[3] found that the lithological composition of the till varied markedly, reflecting local bedrock changes;

> " ... it seems that unconsolidated surficial deposits in the path of the ice have a greater effect on the textural composition than the indurated bedrock, even where the latter is relatively soft shale."[4]

Thus, one must always bear in mind the effects on till composition of the overriding by ice of deposits referable to an earlier glaciation.

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1. Ibid.
2. Charlesworth, J.K., 1959, p. 378.
3. Krumbein, W.C., 1933, p. 392.
4. Op. cit. p. 397.

The colour of a till is a further distinguishing feature, and it naturally varies with the material composing the till.

> "Like the texture and composition, it changes in successive boulder clays gleaned from different sources."[1]

This is well brought out by a description of the Variagated Glacier, Alaska, so named because of its striking multi-coloured moraines:

> "One band is black because of the abundance of black hornblende gneiss; another grey because of the large percentage of white granite boulders; a third purple by reason of the great quantity of gneiss boulders; a fourth is orange-coloured because of a bright orange rust on a gneiss which predominates in it."[2]

Thus, the lithology of the particles in a till reflect the types of bedrock and other materials over which the glacier has passed, and although much of the till is likely to be of local provenance, most contains a proportion of stones different from the rock on which the till lies. These 'erratics' may be defined as

> " ... fragments of rock which ice or its streams have transported from their parent source."[3]

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1. Charlesworth. J.K., 1959, p. 378.
2. Tarr, R.S. & L., Martin, 1906, p. 148.
3. Charlesworth, J.K., 1959, p. 362.

Where till is scanty or absent, erratics may be found lying free on the surface. The exact origin of most erratics is unknown, though the provenance of many can be placed within a specific geologic province. An erratic whose place of origin is known by direct comparison with the bedrock is known as an 'indicator'. To provide good indicator erratics, the parent rock should be sufficiently distinctive to be readily identified and have a limited and well-delineated outcrop.

> "As 'indicator boulders' igneous rocks, notably intrusive types, are generally the more suitable, since the outcrop of sedimentary rocks are usually larger and less well-defined..."[1]

Also, igneous rocks are more durable and thus better able to withstand transport by ice. By plotting the distribution of erratics, particularly indicators, one is able to ascertain the general direction of the flow of the glacier. It should be remembered, however, that movements of ice within an ice-sheet are usually complex, and that more than one glaciation has probably affected the area,

> "... some erratics may have reached their present resting places by devious routes, having been transported in different directions by glaciers at different times."[2]

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1. Op. Cit. p, 368.
2. Sissons, J.B., 1967, p. 72.

## KERRY HILLS

Black Gate col, on the watershed separating the headwaters of the Mule from the Ithon, is covered with considerable deposits of till with a hummocky topography (fig. 3.1.). Dwerryhouse and Miller referred to several moraines within the col[1], but the till is, in reality, dissected by meltwater channels up to 12 ft. deep (pl. 6). The Mule has been 'pushed' against the north-western side of the col and is incised about 30 ft. into the till. This section reveals a heterogeneous, stony deposit with predominantly local erratics, some striated, and some Llandovery grits and shales which are probably derived from the west.

A lower, broader col at Dolfor is also covered with a thick drift deposit. A group of boreholes southwest of the village indicated 28 ft. of clays and gravels[2]. About 4 ft. of brown clay overlies a variable thickness of blue clay, which in one place is separated by 14 ft. of gravels from an underlying blue clay. The upper brown clay may be a weathered equivalent of the blue clay.

A small patch of drift is located on Crygynnau, at 1,570 ft., and to the east, at the head of the Teme, is a broad drift-covered area. The Teme has been ' pushed' against the side of the valley and is incised about 15 ft. into glacial till.

East of the watershed the deep valleys which drain northwards from the Kerry Hills all have deposits of drift within them.

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1. Dwerryhouse, A.R. and A.A. Miller, 1930, p. 99.
2. For full details of all borehole records quoted in this chapter see Appendix 1.

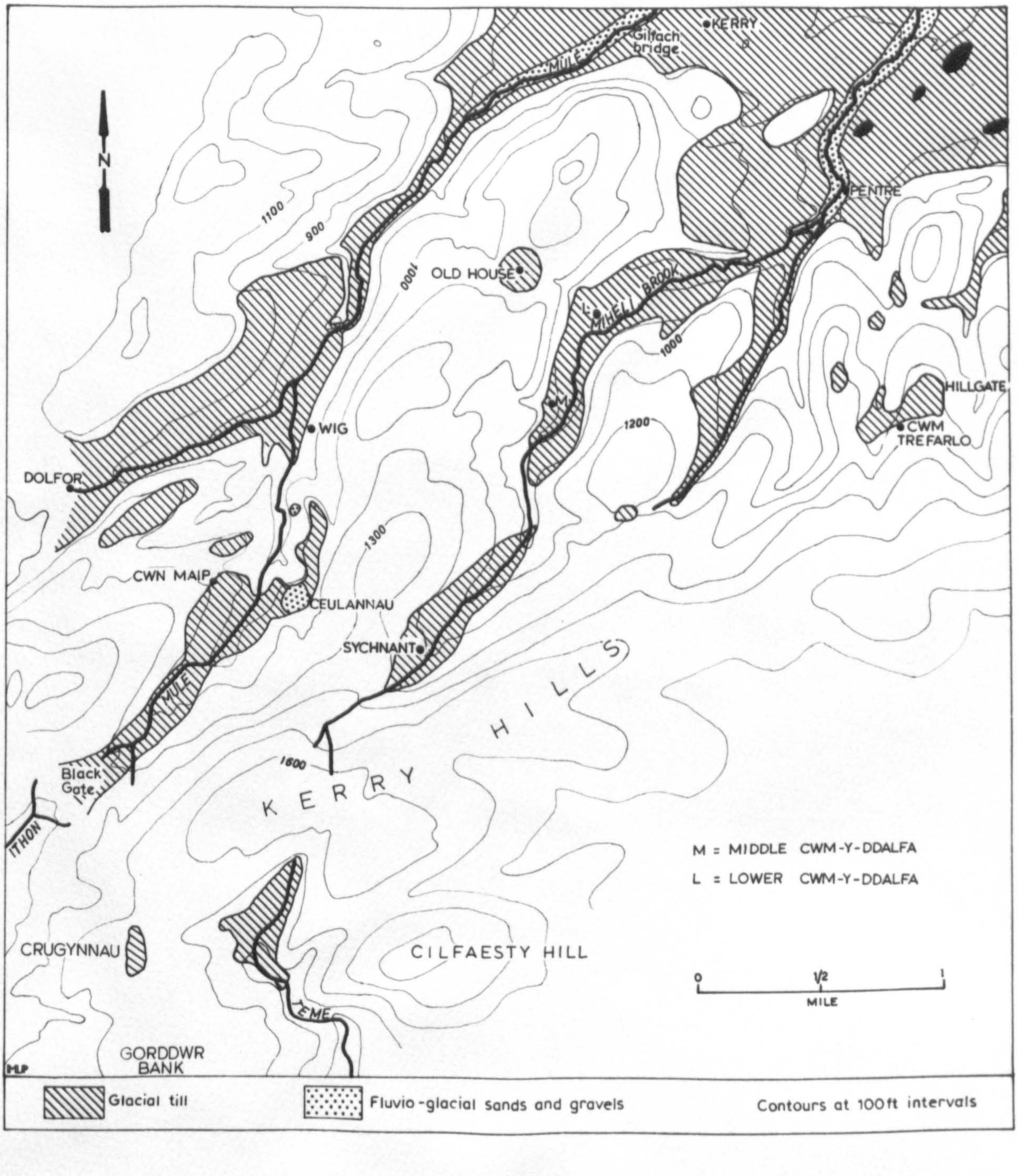


Fig. 3.1 Distribution of drift on the Kerry hills

Pl. 6 Drift in Black Gate col, dissected by meltwater channels

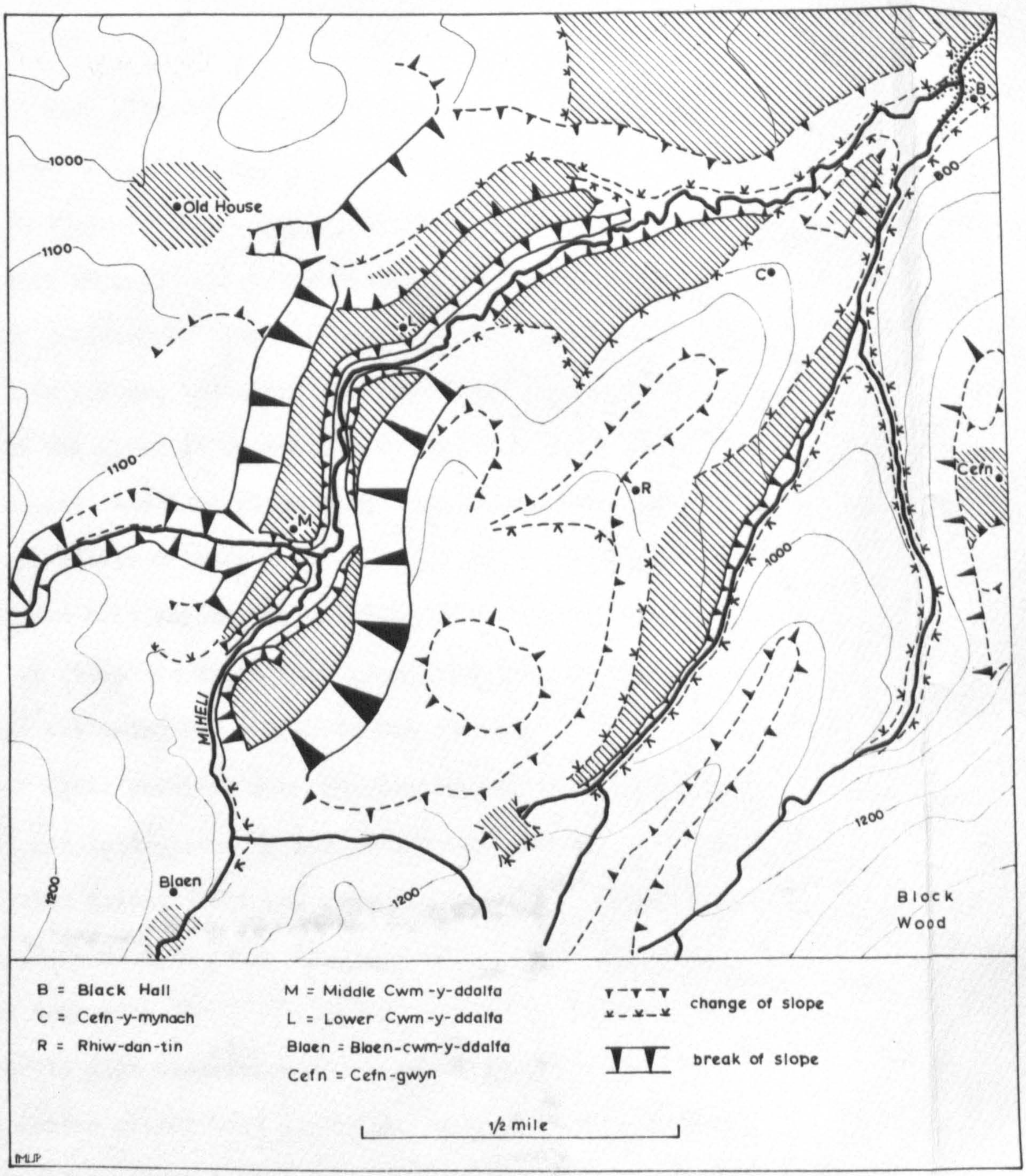


Fig. 3.2 Drift in the Miheli valley. The key for this, and subsequent maps is as in Fig. 3.1

terrace on the west becomes separated from the valley side by a broad, marshy, depression, and terminates as a convex lobe pointing downstream. The two terraces are almost flat-topped; they slope towards the centre of the valley at about 3° above Middle Cwm-y-ddalfa, but below this point they are horizontal. The gap separating the terraces is only 5-10 yd. wide in places, and since they are coincident in height on either side of the river it is considered that they originally formed a continous deposit and have since been dissected by the river. At Middle Cwm-y-ddalfa a large meltwater channel enters the valley from the west and cuts across the terrace. The floor of the channel stands about 10 ft. above the Miheli brook, and has been incised by a small stream following the course of the channel.

Frequent small sections show the terraces to be composed of glacial till, consisting angular and subangular erratics of local origin, Ludlovian grits, flags and shales, in a brown clay matrix. At Lower Cwm-y-ddalfa a 12-foot section in the river bank reveals a complex drift sequence, [illegible]. A lower layer of bedded silt, sands and gravels dips downvalley and is overlain by a lens of well-rounded coarse gravel with no obvious bedding. This in turn is overlain by a brown clay till. The whole deposit passes laterally into a stony till with angular and subangular boulders, predominantly of local grits and shales.

Downstream of the terraces the floor and sides of the Miheli valley are covered by glacial drift, which merges with the broad drift-covered area around Pentre (SO 157888).

The interfluve between the Miheli valley and that to the east is cut through by a col at 1200 ft., the floor of which is drift-covered (fig. 3.1). Downstream, the western side of the valley is masked by a large accumulation of drift which has been cut into by a stream and forms a broad terrace some 20 ft. above the river.

To the east lies the deep, northward-draining valley which carries the main road across the Kerry Hills to the Clun valley. There is a small patch of till in the floor of a shallow col in the interfluve. Till also coveres the upper part of the valley, around Cwm-trefarlo (SO 158875), which extends eastwards up the valley side and onto the interfluve at Hillgate (SO 161878). Other small patches of till occur further downvalley.

A broad col at Shepherd's Lodge, at the head of Drefor dingle, forms part of the divide between the Mule drainage and that of the Caebitra (fig. 3.3). The southern side of the col shows evidence of ice erosion and glacial till covers the floor. The erratics are predominantly of local grits and shales.

Below the col the valley is drift-free; the river is incised into the valley floor and flows in a deep gorge. At Cefn-gwenthrew (SO 190893), however, the valley widens and the river meanders over a broad drift-covered floor. Outcrops in the river banks at Cwm Erl (SO 195898) show about 3 ft. of brown clay with subangular grits and shales overlying bedrock.

Below Cwm Erl, the valley opens out onto the broad Caebitra valley, but the northern side continues as a low, flat-topped spur, for about three-quarters of a mile to a point just south of Sarn

(SO 204908). The spur is a bedrock feature but drift is banked up against its sides, about 5 ft. being exposed at lower Cwm Erl (SO 198904).

The spread of glacial till at the lower end of the Cwm Erl valley is continuous with a similar deposit which extends up to about 800 ft. in the valley to the east. A small patch of till was found on the interfluve at Cwmerl Bank (SO 196894), but generally the upper slopes are drift-free. No further evidence of glacial deposits, in the form of drift or scattered erratics, was seen in this valley or on the interfluves up to the Kerry Hills watershed. Numerous outcrops indicated that about 1 ft. of gravelly head deposits generally overlaid bedrock.

East of Cwm Erl, the next major system of valleys draining the Kerry Hills are those which converge at City (SO 211899). The lower portions of which extends up onto the interfluves. Above about 800 ft. the hillsides and valleys are drift-free, and there are numerous outcrops of head, peat and bedrock.

At Hopton Bank (SO 232896), just north of the Kerry Hills watershed, a small patch of drift occurs adjacent to a system of meltwater channels. The drift is rather thin, and is characterised by a brown clay with local subangular grits, shales and flags. The till contrasts with the nearby 'head' in that the latter is very stony, with angular stones of gravel size set in a gritty matrix. At Argoed Lodge (SO 245904) a broad col at 1075 ft. seperates Pentrenant Dingle from the north-eastward draining Cwmlladron, both of which contain glacial meltwater channels. The floor of the

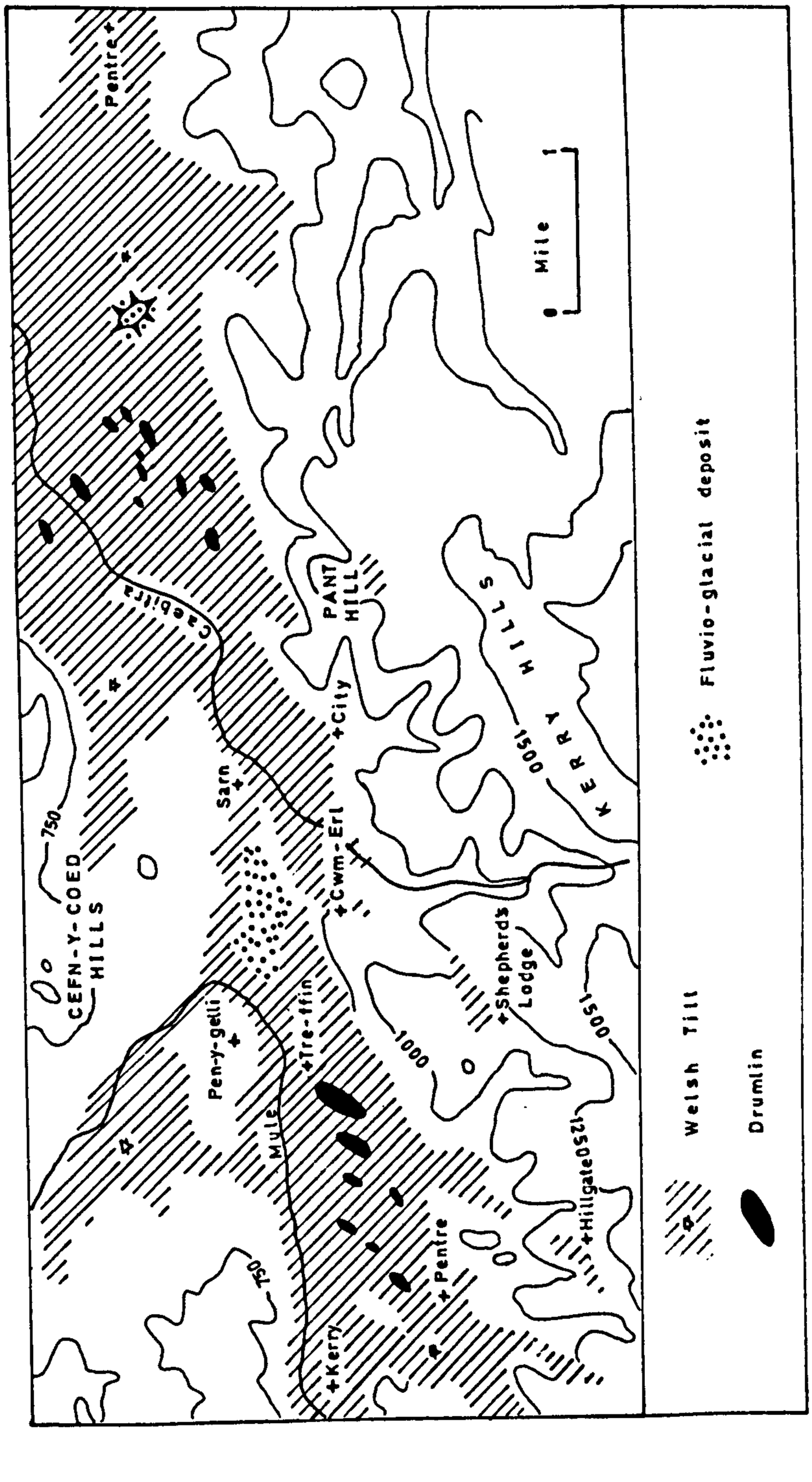


Fig. 3.3 Drift on the eastern Kerry Hills and Mule - Caebitra valley

col is covered by glacial till.

The lower part of Cwmlladron is covered by a thick deposit of till. A number of sections to the southwest of Cwm (SO 259909) reveal 5-6 ft. of till, including erratics of grit and shale; the latter are heterogeneous in size, ranging from small gravel-sized pebbles to large boulders 1-2 ft. across. This till covers the hillsides around Cwm and the area to the east up to about 800 ft; it is patchily distributed above this height, right up to the watershed at 1250 ft. In these higher areas, scattered erratics are also to be found; these are generally large blocks of grit and shale, occasionally striated, and occasional quartzites.

## THE MULE - CAEBITRA VALLEY

Between the steep front of the Kerry Hills and the Cefn-y-Coed hills to the north is a broad depression traversed by the rivers Mule and Caebitra. Much of the depression is covered by thick and varied deposits of glacial drift (fig. 3.4). A borehole 250 yd. northeast of Gilfach Bridge (SO 141900), to the west of Kerry, indicated the following sequence of deposits:

| | Thickness | Depth | (ft.) |
|---|---|---|---|
| Topsoil - loam | 2-0 | 2-0 | |
| Loam - clay | 5-0 | 7-0 | |
| Fine gravel | 9-0 | 16-0 | |
| Sandy clay | 7-0 | 23-0 | |
| Clay | 13-0 | 36-0 | |
| Alluvial grit and sand | 64-0 | 100-0 | |

This suggests that a considerable thickness of glacial and fluvio-glacial deposits occurs in this locality

Southeast of Kerry is a broad embayment in the Kerry Hills, at the confluence of the Miheli brook and its tributaries. The valley floor and hillsides, up to about 700-750 ft., are covered by a gravelly drift deposit. At Pentre a broad, drift-covered terrace, whose surface is at 727 ft., is situated at the foot of the Kerry Hills; a similar feature on the opposite side of the river has bedrock exposures on its surface. At Drefor, about a mile east of Pentre, is another drift-covered terrace at a similar altitude.

Perhaps the most impressive features in this area is the group of drumlins to the northeast of Pentre, the long axes of which are orientated northeast to southwest, directly in line with the upper Miheli valley. They are up to 30 ft. in height. They are not all composed entirely of drift however, since bedrock exposures occur in the surface of the drumlin at Pen-y-bryn (SO 177901).

Although much of the area around Kerry is covered by till deposits, fluvio-glacial gravels are consistently exposed in the river banks. Sections in the Mule show 1-2 ft. of alluvial material overlying well-sorted, bedded gravels with occasional layers of silt. Similar material is exposed along the banks of the Miheli and Drefor brooks, and extends back into the deep, steep-sided valleys of the Kerry Hills. At Blackhall Wood, to the south of Pentre, a section showed 6 ft. of glacial till overlain by 2 ft. of a gravelly fluvio-glacial deposit. At Glan-Mule (SO 162904), the Miheli brook has cut down about 10 ft. into a fluvio-glacial deposit (Pl. 7). Large

Pl. 7 Section in fluvioglacial outwash material in the Miheli Brook valley

Pl. 8 Till deposit at Goetre Brick Works

sub-rounded boulders, up to 1 ft. across, are set in a matrix of well-rounded gravels; the larger boulders are flattened and plate-shaped. Interbedded in this deposit were two lenses, up to 1 ft. thick, of reddish-brown sand. The erratics included in the deposit are predominantly local Silurian grits, sandstones, shales, and flags. The gravels are most probably outwash deposits, laid down over the till by meltwaters from the retreating ice.

To the east of Glan-Mule the Mule continues to flow across a broad drift-covered valley floor. Numerous sections away from the river show a blue clay till with included grits and shales, generally sub-angular and heterogeneous in size. The drift is continuous through the gap to the west of Pen-y-Gelli hill.

A borehole to the southwest of Pen-y-gelli (SO 176905), adjacent to the river, revealed the following sequence of deposits:

| | Thickness | Depth | (ft) |
|---|---|---|---|
| Brown soil | 4-0 | 4-0 | |
| Gravel | 2-0 | 6-0 | |
| Blue clay | 69-0 | 75-0 | |
| Gravel | 29-0 | 104-0 | |

The brown soil is probably alluvium and the underlying gravel an outwash deposit; the underlying blue clay and gravel are true glacial deposits. At Pen-y-gelli, as at Kerry, therefore, over 100 ft. of deposits fill the valley, the bedrock floor of which is at an unknown depth.

About half a mile east of Glan-Mule, on the southern side of the river, is a conspicuous isolated kame, about 15 ft. high,

a small section in which showed dirty, poorly sorted gravels. A group of similar kames occurs 700 yd. further east at Tre-ffin (SO 180902), where the Drefor brook joins the Mule. Here,the Drefor and Mule have cut a terrace about 15 ft. high, composed of bedded fluvio-glacial outwash gravels.

The two kame deposits are outliers of a more prominent and complex area of drift topography immediately east of the point where the river Mule makes a sharp turn towards the northwest. The drift deposits reach up to about 600 ft. and extend right across the valley. Adjacent to the Mule is a complex series of irregular kames and ridges up to 25 ft. high, with occasional marshy kettle holes, 5-10 ft. deep. Farther east, the topography becomes more subdued and undulating, with isolated kames. Sections within the kames and ridges reveal well-rounded, water-worn gravels in a sandy-loam matrix.

Eastwards, the kame-complex merges into a relatively flat drift-covered area at 600 ft., which extends past Llwyncowrid (SO 196909) to Sarn (SO 204908). The existing watershed between the drainage of the Mule and the Caebitra is near Sarn Church, at a height of 610 ft. Some indication of the thickness of drift deposits between the Mule and Sarn is given by several borehole records. At Llwyncowrid a borehole revealed at least 50 ft. of drift; to the east, at Sarn, three bores[1] indicate that about 31 ft. of glacial till overlies bedrock. The height of the bedrock floor between the Mule and the Caebitra therefore lies below

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1. Full details of the borehole records are given in Appendix 1.

550 ft. at Llwyncowrid and at about 569 ft. to the east of Sarn.

Beyond the kame complex, the Mule flows first northwestwards in a wide, drift-covered valley east of Pen-y-gelli hill; it then turns north to flow through a deep, winding, narrow gorge to Abermule (SO 163945), where it enters the Severn.

At Pen-y-gelli Mill (SO 186914) a broad terrace at 500 ft. is situated adjacent to the river. Exposures show it to be a bedrock feature, but the surface is covered by till. On the opposite side of the river a terrace at the same height is formed entirely of till, which continues over the slopes of Pen-y-gelli hill. The two terraces can be traced downstream as far as Goetre (SO 184918), and the hillsides are mantled with drift up to about 800 ft.

At Goetre Mill (SO 175922), a considerable thickness of glacial till has been deposited in the valley floor, and is at present being quarried for a brickworks. A 20 ft. section in the quarry face (PL.8) showed up to 5 ft. of stiff brown clay overlying 15 ft. of very stiff blue-grey clay. Both clays have included subangular pebbles and boulders and these are predominantly Silurian mudstones sandstones and shales, with some quartzite pebbles. The two clays appear to be part of the same deposit, and the brown colouration of the upper clay may be due to weathering. The exact thickness of this deposit is not known.

Till of a similar character extends downstream from this point and forms a broad terrace on the western side of the river. At Lower Maenllwyd (SO 165928) the river enters its gorge section in

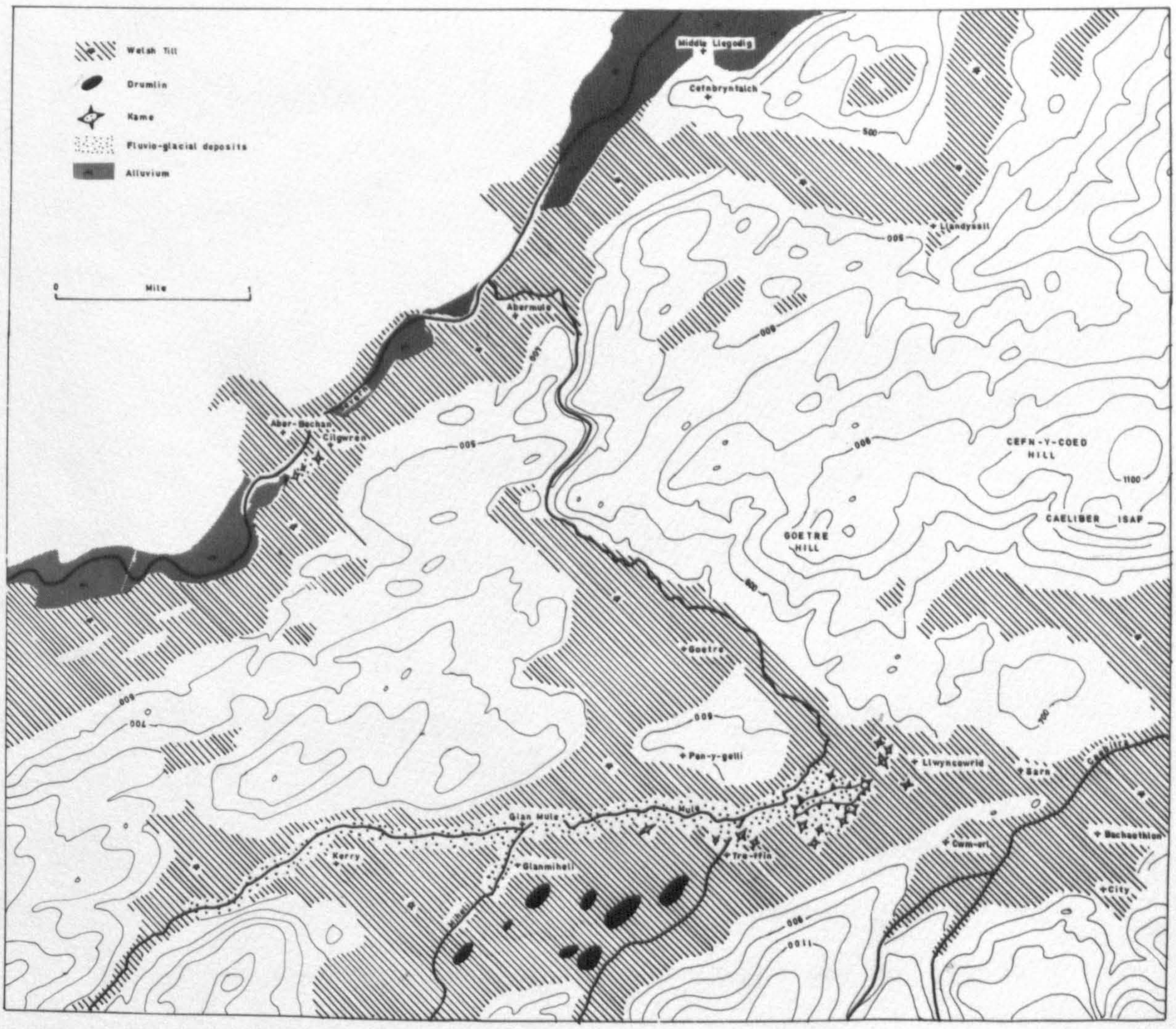


Fig. 3.4 Drift in the Mule and Severn valleys

which no deposits of glacial drift were recorded.

Southeast of Sarn in a broad undulating plateau at about 600 ft. lies on the southern edge of the Caebitre valley (fig. 3.5). The area is covered by gravelly till, which extends up to about 700-800 ft. on the lower slopes of Kerry Hills. Rising out of the plateau are a series of drumlins, up to 15 ft. high, the long axes of which are orientated roughly northeast-southwest. They are covered by till, but since few exposures were found it is not possible to tell whether any have a solid core.

Further east, extensive deposits of drift cover the lower slopes of the Kerry Hills south of Mellington Hall (SO 258920). At many localities up to 6 ft. of till, comprising a brown clay with included sub-rounded and rounded erratics of various sizes, was observed.

On the northern side of the valley is a similar covering of till. It is patchily distributed over the hills immediately north of Sarn, and covers the floor of the deep, steep-sided valley between it and Caeliber Isaf. Up to 12 ft. is exposed in the valley floor west of Gwmberllan (SO 202925). Drift is not found on the steep slopes of Caeliber Isaf, but patches occur up to 800 ft. on Caeliber Uchaf and near Great Cefnyberin to the west.

At Church Stoke, a section by the Methodist Chapel revealed 4 ft. of stiff blue-grey clay, with included sub-rounded erratics of sandstones, grits, shales, and mudstones. A small section in the garage to the east showed 10 ft. of similar blue-grey clay, with

included boulders up to 10 in. across; a boulder of locally-derived fine-grained igneous rock measured 2½ ft. across. Small sections revealed a similar deposit extending westwards around the brickyard claypits, eastwards beyond the village towards the lower slopes of Todleth Hill, and northwards up the gorge of Marrington Dingle.

West of Church Stoke, the broad floor of the Caebitra valley is covered by stony glacial till which forms a series of low irregular undulations with small enclosed depressions. The latter are up to 30 ft. across and 10 ft. deep; many are marshy and some have water in them. They are interpreted as a series of small kettle holes.

Immediately to the north, and extending towards Montgomery is a group of steep-sided, irregular kames and ridges, up to 30 ft. high. These are especially well-developed to the east of Llwynobin farm (SO 229954) and at Gwarthlow (SO 246954). A section in a kame near Parkside (SO 235960) showed well-stratified stony drift. The sub-rounded erratics range in size from boulders 2½ ft. across to fine sand; the predominant size-range being ½-2½ in. and there is very little clay. The deposit is segregated into bands and lenses, each showing a general uniformity of grain-size. The larger fragments are entirely of shale, and the others are mudstones, sandstones and grits, with occasional rounded pebbles and boulders of basic igenous material.

Alluvium and bedded fluvio-glacial outwash deposits occur adjacent to the river Caebitra. Sections in the river banks west

of Old Hall (SO 205903) show 6 ft. of bedded gravels. The pebbles are predominantly rounded and are of heterogeneous size and ill-sorted; some are angular and fairly large, up to 1 ft. across. This deposit can be traced up stream to Wern, at which point the river bifurcates. Here, bank-side sections show well-bedded gravels with abundant small rounded pebbles. In places the gravels overlie a stiff blue-grey clay, which is probably till, and in others up to 1ft. of brown loamy clay, probably alluvium, overlies them. Further upstream, at Brompton Bridge (SO 251931), are extensive deposits of fluvio-glacial gravels.

## CEFN-Y-COED HILLS

The Cef-y-Coed Hills reach up to a maximum of 1158 ft. in the west, and are severed in two by the valley of the Mule. The area is covered by glacial drift up to about 700 ft. above which the drift-cover is patchy and confined to the floors of valleys and depressions (fig. 3.4).

In the southeast, by Newtown, the area adjacent to the Severn valley is covered by till up to about 500 ft. This is a brown gravelly clay with boulders, up to 1 ft. across, of Silurian grits and shales; these are generally sub-rounded and a few are striated. Generally, the till cover is only thin, between 3 and 5 ft. resting directly on bedrock, and is concentrated in the linear depressions between the ice-moulded ridges.

At Llandyssil (SO 197953) a broad, steep-sided valley opens out at both ends onto the valley of the Severn, and severs from the mass of high ground to the south a hill whose gently undulating surface is at 625 ft. There is a low watershed in the centre of the valley at 386 ft. The floor is covered by an unknown thickness of glacial till, a brown gravelly clay with included grits, flags and shales. A borehole at Llandyssil, adjacent to the Church, indicated about 38 ft. of till resting on bedrock, the surface of which must be at about 361 ft. It seems probable, however, that the deposits thicken towards the centre of the valley and the bedrock floor may lie below 350 ft. In the centre of the valley, adjacent to the streams, is a deposit of roughly-bedded waterworn gravels. At the northern end of the valley, east of Rhyd-y-Ware (SO 199969) a gravel pit in the side of a small kame shows unbedded and poorly-sorted sands and gravels.

## MONTGOMERY-CHIRBURY GAP

Between the Cefn-y-Coed Hills and the high Ordovician country to the east, dominated by Corndon Hill, lies a broad depression underlain by Wenlock Shales. This forms a divide between the Caebitra-Camlad valley on the south and the Marton-Camlad valley to the north, the two being connected by Marrington Dingle which is cut across the divide (fig. 3.5). The depression is highest in the south, where a steep edge overlooking the Caebitra valley

reaches up to 500 ft. Much of the area forms a gently undulating plateau between 400 to 450 ft. and is dissected by a number of valleys which head northeastwards to a stream heading north from Sidnal (SO 263964) towards Chirbury (SO 262983) and the Camlad valley.

The depression is covered by a variable thickness of glacial and fluvio-glacial deposits. The till is typically a stony brown clay with erratics of local Silurian shales, and grits. These are of variable size, generally sub-rounded to angular, and a few are striated. In the west, in Lymore Park, the drift takes the form of gently undulating topography, but elsewhere it forms a thin cover over the plateau surface. That bedrock is at no great depth is shown by a borehole record at Rhiston, (SO 259951) which indicated only 2 ft. of superficial material.

At Moat (SO 247968) and at Timberth (SO 252965), both of which are situated on the interfluves between the northeast-draining valleys, only 7 to 8 ft. of clay, glacial till, overlies the bedrock, the surface of which is at about 425 ft. There is an outcrop of bedrock to the northeast of Timberth, and there are numerous small shale quarries, known locally as 'marl pits', on the tops of interfluves and hillocks. These occur, for example, at Walcot (SO 264994), northeast of Rownal (SO 234983), near Winsbury (SO 247983), and above Marrington Dingle to the north of Heightly (SO 271988).

The cover of glacial drift is generally thicker in the valleys and depressions. The principal northward-flowing stream is cutting

into the drift by as much as 20 ft. at Chirbury, and an exposure to the north of Chirbury reservoir revealed 10 ft. of till deposits. That a considerable thickness of glacial material has been deposited around Chirbury is also shown by a borehole at The Works (SO 269983), in the valley floor, which revealed up to 38 ft. of gravels and clays resting on bedrock.

On the northern edge of the plateau, near Winsbury (SO 247983) and Dudston (SO 245975), are several prominent hillocks in an area of undulating drift topography. Small sections show a high proportion of rounded gravels in a sandy matrix and it is considered that the hillocks form part of a kame,- similar to that on the southern edge of the plateau.

Other small patches of fluvio-glacial deposits occur adjacent to the Camlad valley. At Walcot farm a quarry in the side of the valley heading north across the plateau from Chirbury shows 10 ft. of gravels resting on a flat surface of Wenlock Shale. The gravels are roughly-bedded and poorly sorted. The lower 4 ft. consist of well-rounded pebbles, up to 3 in. across, in a sandy matrix; a lens of sand 1 ft. thick and 6 ft. long occurs within this lower gravel. The upper 6 ft. is generally coarser, with boulders up to 8 in. across.

East of Marrington Dingle is an irregularly dissected area between 500 to 800 ft., backed by the steep slopes of Todleth, Llan Fawr and Corndon. Except for the summit of the prominent northeast to southwest aligned Ridge, the whole area is covered by

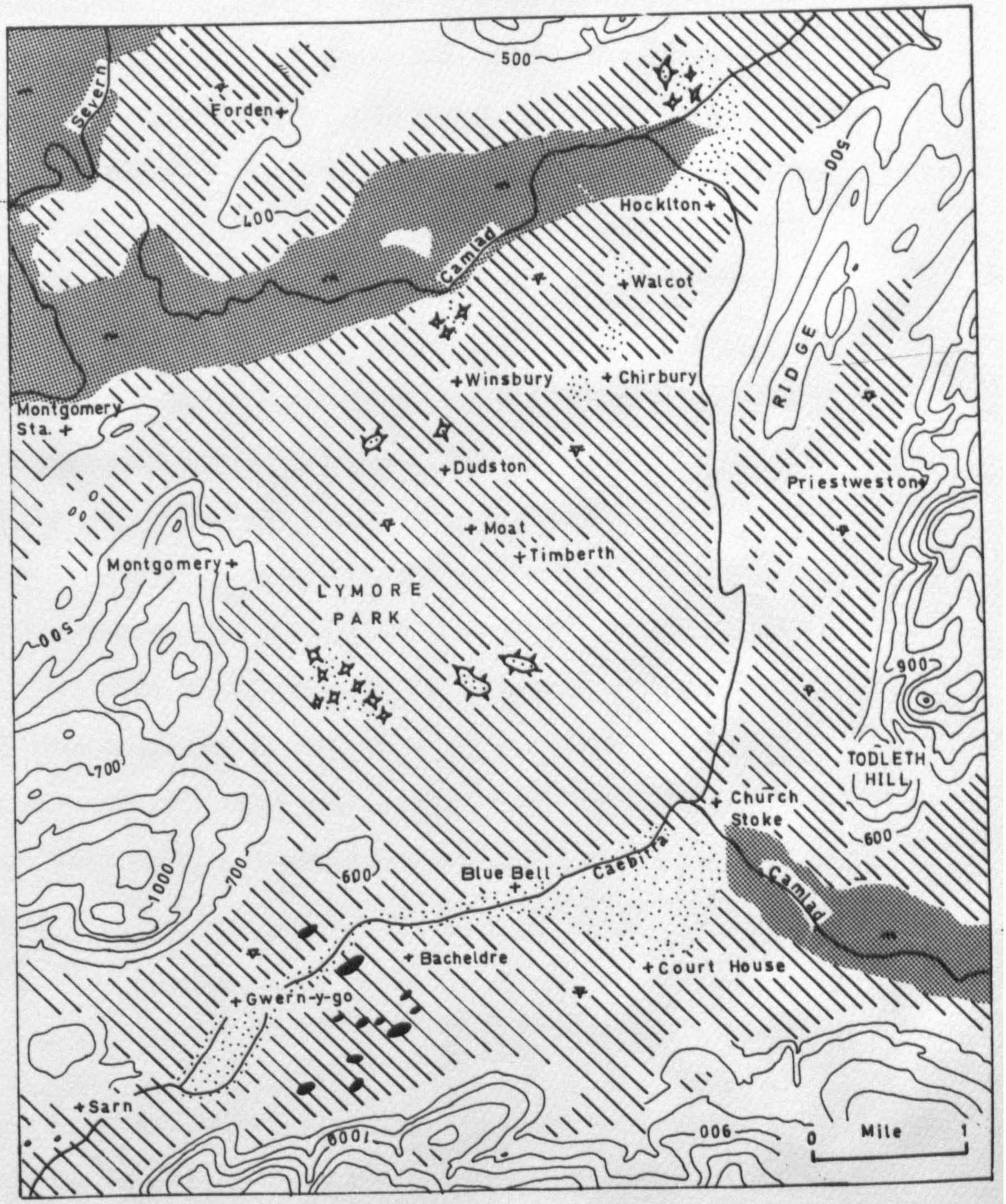


Fig. 3.5 Drift in the Caebitra valley and Montgomery - Chirbury gap

glacial deposits. These are characterised by a brown clay, with included erratics of Silurian grits and shales and a substantial proportion of local Ordovician volcanics. The erratics are generally rounded to sub-rounded, varying in size from gravels to boulders, and a few bear striations. A number of wells west of Old Church Stoke (SO 296949) record boulder clay up to 12 ft. in thickness overlying bedrock.

The steep hillslopes which form the eastern margin of this area are generally devoid of drift, except for a small pocket of till in the col at The Pant (at 800 ft.) above, Old Church Stoke.

The sides of Marrington Dingle are entirely in bedrock, but there are some traces of till in the floor of the gorge. A section at Whittery Bridge (SO 272983), east of Chirbury, showed 1 ft. of alluvial clay overlying a heterogeneous stony clay. A similar deposit was exposed further south, near Calcot (SO 273959), while at Alport (SO 274953) a borehole revealed 8 ft. of clay resting on shales.

## REA BROOK - CAMLAD VALLEY

The broad Rea Brook - Camlad valley heads northeastwards away from the Severn at Montgomery, separating the Long Mountain from the hills of the Corndon District (fig. 3.6). Along the northern side of the valley an undulating platform at 450-600 ft. is backed by the steep edge of the Long Mountain. Superimposed on the platform are a series of sub-parallel bedrock ridges of

drumlinoid form. The area is covered by glacial drift, which is generally only thin over the ridges, but thicker in the minor valleys and depressions. The till is a stony, buff-coloured clay with included subangular shale fragments up to 6 in. across, sandstones, grits, and occasional volcanic ash. At Aston Rogers (SO 318045) a stream ~~is~~ has cut~~ting~~ down some 3-5 ft. into this till; while a similar deposit is to be found at Brockton (SO 318045), Binweston (SO 302042) and north of Little Worthen (SO 333052). A well record at Aston Piggott (SJ 059338) showed 8 ft. of 'brown stony marl', probably a glacial till, overlying bedrock.

The southern side of the valley has no equivalent platform, and the steep hillslopes are largely devoid of drift. Small patches do occur however, in depressions and on flatter areas, as for example at Wilmington (SJ 297091), Betton (SJ 317027) and Grimmer Farm (SJ 344039).

The most prominent accumulation of drift is in the floor of the valley. At the southwestern end, adjacent to the Severn, the river Camlad flows over a broad alluvial flood plain. Numerous bankside sections reveal 3-5ft. of a dark-grey stoneless silt, with occasional lenses of gravel. That this is underlain by glacial deposits is shown by a borehole record at Gaer Bridge (SO 214999), on the northern side of the valley, which revealed 6 ft. of alluvium overlying 21 ft. of stony clay which, in turn, rested on bedrock. Immediately upstream from the point at which the Camlad emerges from Marrington Dingle and makes a sharp turn towards the southeast

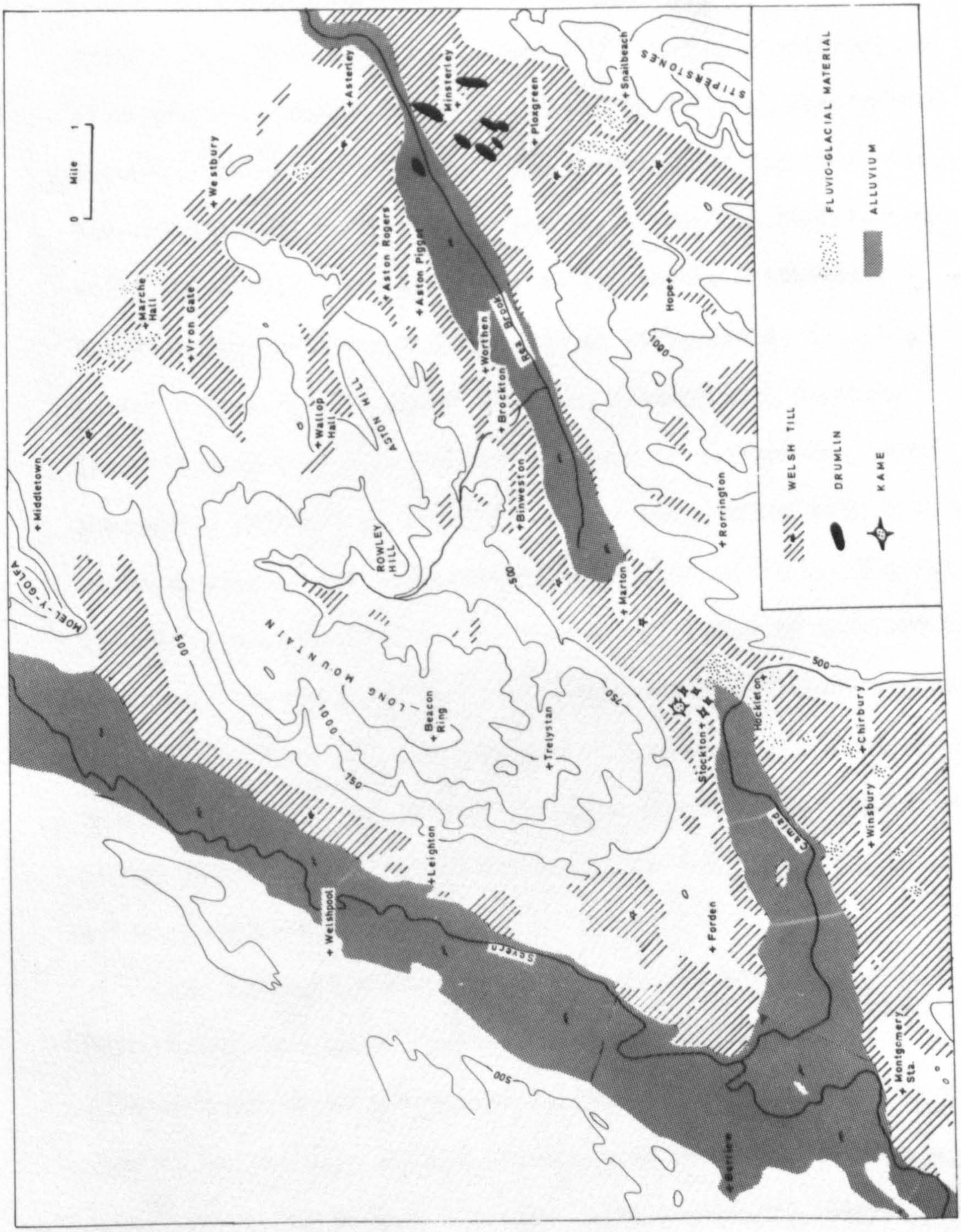


Fig. 3.6 Drift on Long Mountain and in the Rea Brook - Camlad valley

there is a prominent terrace feature extending across the valley (fig. 3.7). This has a steep face (20°) facing downstream and rises abruptly some 25 ft. above the flat alluvial flood-plain extending towards the Severn. The terrace edge separates the low-lying Camlad valley to the southwest from the higher, but still relatively flat, Rea-Brook valley to the northeast. A broad embayment in the terrace edge at Cross Lanes, (Pl. 9), is cut into by a deep gully revealing a stony till deposit. The till comprises a pale-brown clay and is ill-sorted, large subangular boulders up to 2½ ft. across being mixed with a variety of fragments and particle sizes down to clay and silt. The smaller fragments, of gravel size, are generally rounded to sub-rounded in shape. The most abundant constituents are Paleozoic shales, slates and grits, but there were also specimens of sandstone, quartzites, and a few of volcanic ash. The deposit is poorly-bedded and a lens of sand 2 in. thick and 6 ft. long gives some evidence of stratification.

The surface of the terrace is relatively flat, but is diversified by a small number of kettle holes 10-15 ft. deep, west of Cross Lanes and around Aylesford Bridge. On the northern side of the valley, towards Stockton the terrace undergoes a transition to an area of hummocky drift topography. Kames, 20-25 ft. high, occur both on the terrace and adjacent to it; a number of small sections show these to be composed of rather dirty sands and gravels.

Pl. 9 Edge of fluvioglacial terrace in the Marton valley, opposite the Marrington Dingle outlet.

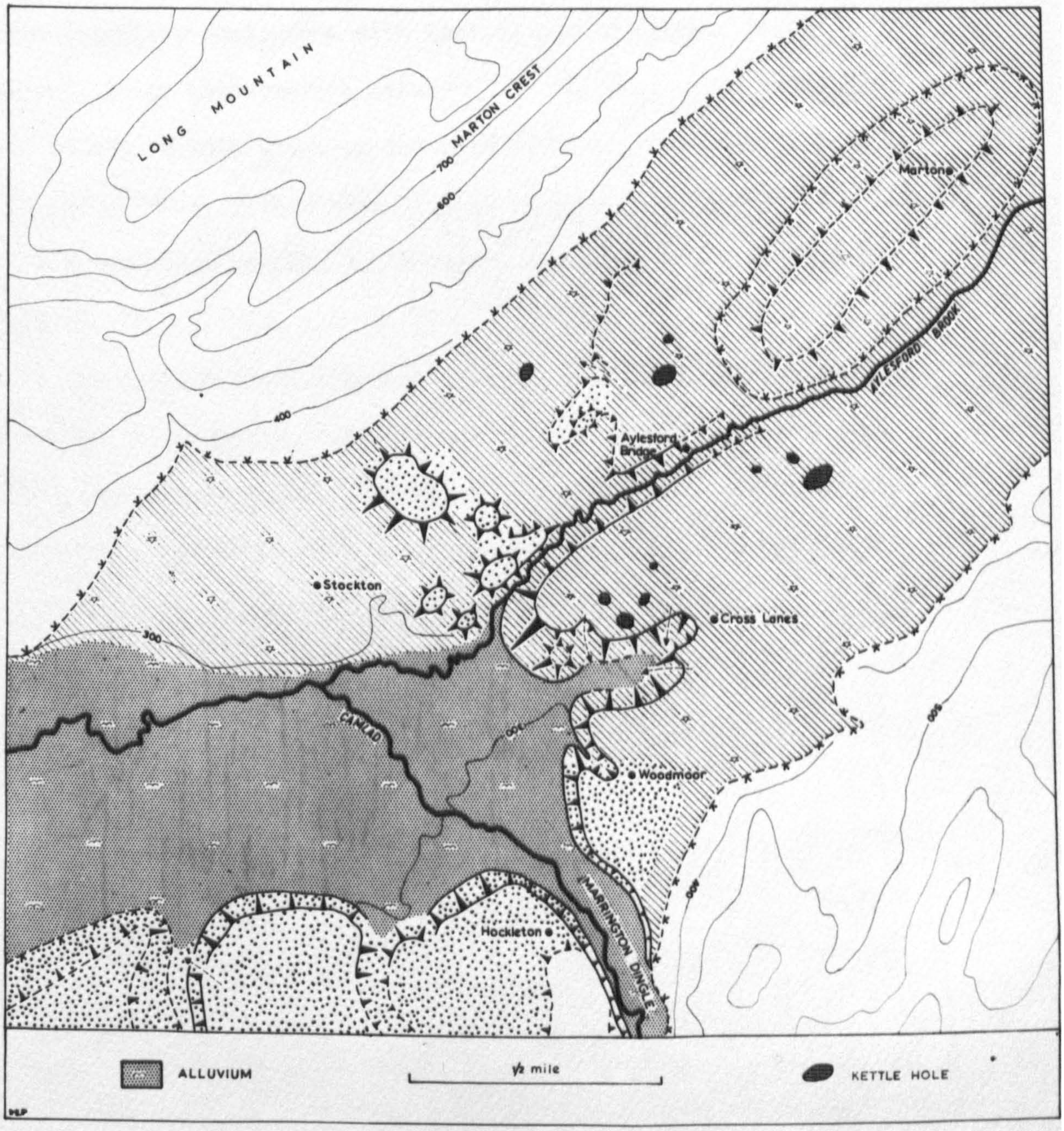


Fig. 3.7 Drift in the Wamlad valley, opposite Marrington dingle

At its exit at Hockleton, Marrington Dingle is incised some 15-20 ft. into a flat-topped terrace feature, which is morphologically continuous with that described above. Here, however, there are numerous exposures of fluvio-glacial sands and gravels on both sides of the gorge, showing that the terrace is composed of this, rather than till as at Cross Lanes. At one point, close to the southern hills, the gravels were observed overlying bedrock.

The relatively flat terrace surface continues towards Marton, the ridge on which the village stands rising prominantly above it, and beyond towards Westbury at the northeastern end of the valley. The valley floor north of Marton ridge is covered by glacial deposits, which extend up to about 400 ft. against the Long Mountain (fig. 3.6). Till is also exposed along the northern side of Marton Pool, and a trench section nearby showed a stony brown clay with shale fragments up to 3 in. across.

The area around Marton Pool is covered with peat and alluvial deposits. The presence of drift below the alluvium is proved by a borehole sunk in the middle of the valley opposite Binweston Farm (SJ 308035):

| | Thickness | Depth | (ft.) |
|---|---|---|---|
| Late glacial gravels | 22-6 | 22-6 | |
| Grey clay with many stones | 9-0 | 31-6 | |
| Coarse sand grains, mainly of shale | 0-9 | 32-3 | |
| Grey clay with many stones | 2-9 | 35-D | |
| Soft grey clay with layers of fine grit | 13-0 | 48-0 | |

| | | |
|---|---|---|
| Soft grey clay with silt and a few stones | 2-0 | 50-0 |
| Hard grey clay with abundant stones | 20-0 | 70-0 |
| Wenlock Shale | 70-0+ | |

This borehole reveals a sequence of till and stratified drift underlying 22½ ft. of Late and Post-Glacial gravels, clays and peat. Since the surface of the deposits is at about 350 ft., the bedrock floor of the valley must be at about 280 ft.

Northeast of Marton Pool, towards Brockton, sections showed that much of the valley floor is covered by peat, or alluvium underlain by peat. Although the peat dies out at Brockton, the alluvial deposits continue towards Lower Hogstow (SJ 357051) beyond which they are underlain by a layer of gravels of angular and subangular shales with no well-marked bedding.

Beyond Lower Hogstow the hitherto flat valley floor is diversified by a number of low drift mounds, for example, at Hem (SJ 356063), west of Westly (SJ 364070), and south of Asterly (SJ 373070). They are generally of irregular form, rise about 25 ft. above the valley floor, and adjacent to the one west of Westley is an enclosed kettle-like depression some 8 ft. deep. Similar drift-mounds occur on the southern side of the valley around Minsterley (SJ 375061). They have a distinct drumlinoid form, are up to 30 ft. high, and their long axes are orientated roughly NNE-SSW. A borehole at Minsterley Creameries indicated that 19 ft. of Made ground and alluvium overlies 17 ft. of boulder clay resting on bedrock.

Scattered over the fields between Minsterley and Asterley and around Westbury are occasional distinctive erratics of dark-green dolerite, probably carried from the Breiddens, and some specimens of Carboniferous Limestone.

The erratics indicate a movement of ice from northwest to southeast in this area, and the Carboniferous Limestone is probably derived from the outcrop at Llanymynech, southwest of Oswestry. Breidden Hill dolerites do not occur as erratics west of the Long Mountain or in the basins of the Caebitra and Camlad, but may be traced southeast to Church Pulverbatch on the northern slopes of the Long Mynd.

The ground east of Minsterley is covered with glacial drift, which is banked up against the western slopes of Callow Hill and the Stiperstones. Small sections showed a brown loamy gravelly clay with numerous subangular and angular erratics, mainly of local origin. In the floor of a meltwater channel north of Polesgate (SJ 392052) 12 ft. of glacial drift was exposed. About 9 ft. of brown loamy gravelly clay was underlain by at least 3 ft. of a tough compacted blue clay with numerous well-striated boulders, some of considerable size. A further example of a blue clay underlying a stony brown clay till is given in a borehole record at Whitton Farm (SJ 344098), where 50 ft. of glacial drift overlies bedrock.

## LONG MOUNTAIN

That the Long Mountain has been invaded, at some time, by glacier ice is indicated by the presence of glacial deposits and meltwater channels. The deposits do not form a continuous cover over the surface, the steeper slopes being characteristically drift free, but they occur in patches, often in cols and depressions, right up to the watershed at 1325 ft.

East of the watershed, at Becon Ring (SJ 265058), glacial deposits are exposed in numerous small sections at the head of the Mulsop valley and on adjacent hillslopes. These show a stony brown clay with included subangular and sub rounded erratics of varying sizes from gravels to cobbles and boulders. The erratics are predominantly local shales and mudstones. Further downvalley, at its junction with Rowley Brook, the Mulsop stream has cut down some 10 ft. into the till. Similar deposits are found in the floors of the valleys draining eastwards from Welsh Harp (SJ 279080). Drift is also common on the hillslopes and valley floors around Rowley (SJ 306065). A section on the slopes of Rowley Hill showed 8 ft. of brown loamy clay with included erratics of local origin. There are scattered deposits of drift on Heldre Hill, generally located in valley bottoms. A borehole at The Hill farm (SJ 308100) showed 6 ft. of clay overlying bedrock.

To the southeast, several sections in the flat floor of the Wallop valley show a brown stony clay till. A borehole at Lower

Wallop (SJ 324076) went through 90 ft. of 'stony marl' and gravels before reaching bedrock, suggesting a considerable thickness of glacial deposits in this locality. Similar deposits can be traced down the Wallop valley to the entrance of a gorge at Causebeech, in which it is absent. Beyond the gorge the drift continues beyond Whitton Hall (SJ 245091) to Vennington (SJ 339097) and covers the hillslopes to about 600 ft. In this region the drift contains a increasingly large proportion of rounded pebbles and the matrix becomes progressively sandier.

## SEVERN VALLEY

Between Newtown and Abermule (SO 163946) the southeastern side of the Severn valley is covered with thick deposits of glacial drift. In several places prominent terraces have been formed and in others there are conspicuous moraine-like mounds (fig. 3.4).

At Gro-Cottages (SO 125921) there is a terrace some 20 ft. high which continues downstream for half a mile to just beyond Glan Hafren (SO 131923).

At Penarth (SO 140924), there are two terraces, each some 12 ft. high, which terminate in an area of hummocky drift. Sections in the terraces show a loamy clay with varied subangular and subrounded gravels, generally of local origin. On the opposite side of the valley, above Dolfor Farm (SO 135926) is a terrace some 25 ft. high with flat surface at 325 ft.; by contrast this is a bedrock terrace with a only thin covering of drift.

At Aber Bechan (SO 143935), is an area of hummocky morainic topography, and small sections reveal well-rounded fluvio-glacial gravels set in a sandy loam matrix. A river terrace some 12 ft. high leads away from the mounds towards Cilgwrgan farm (SO 147935).

At Abermule a series of boreholes demonstrate a marked contrast in the thickness of drift on opposite sides of the Severn valley (fig. 3.8). Adjacent to the station, on the south-eastern side of the valley, a maximum of 47 ft. of marls and gravels has been recorded; at The Court (SO 159994) 29 ft. of marls, sands and gravels overlie bedrock. About 500 yd. downstream of Abermule, at Bryn Derwen Bridge (SO 162952), five boreholes have been put down between the main road and the river. On the north-western side of the valley only 6-7 ft. of brown clay and gravel rest on bedrock; towards the opposite side of the valley there is at least 36 ft. of glacial deposits. At Abermule Mills (SO 161949), on the north-western side of the valley, an abandoned meander of the Severn has a bedrock floor, and between The Court and Bryn Derwen (SO 165953) the river is flowing on a bedrock floor.

There is thus a considerable thickness of deposits on the south-eastern side of the Severn at Abermule, while bedrock is close to the surface on the opposite side of the valley. Since the present valley floor is flat, it would appear that a buried channel on the south-eastern side has been filled with glacial drift.

Downstream from Abermule the Severn flows over a flat flood plain, and numerous sections consistently show about 6 ft. of

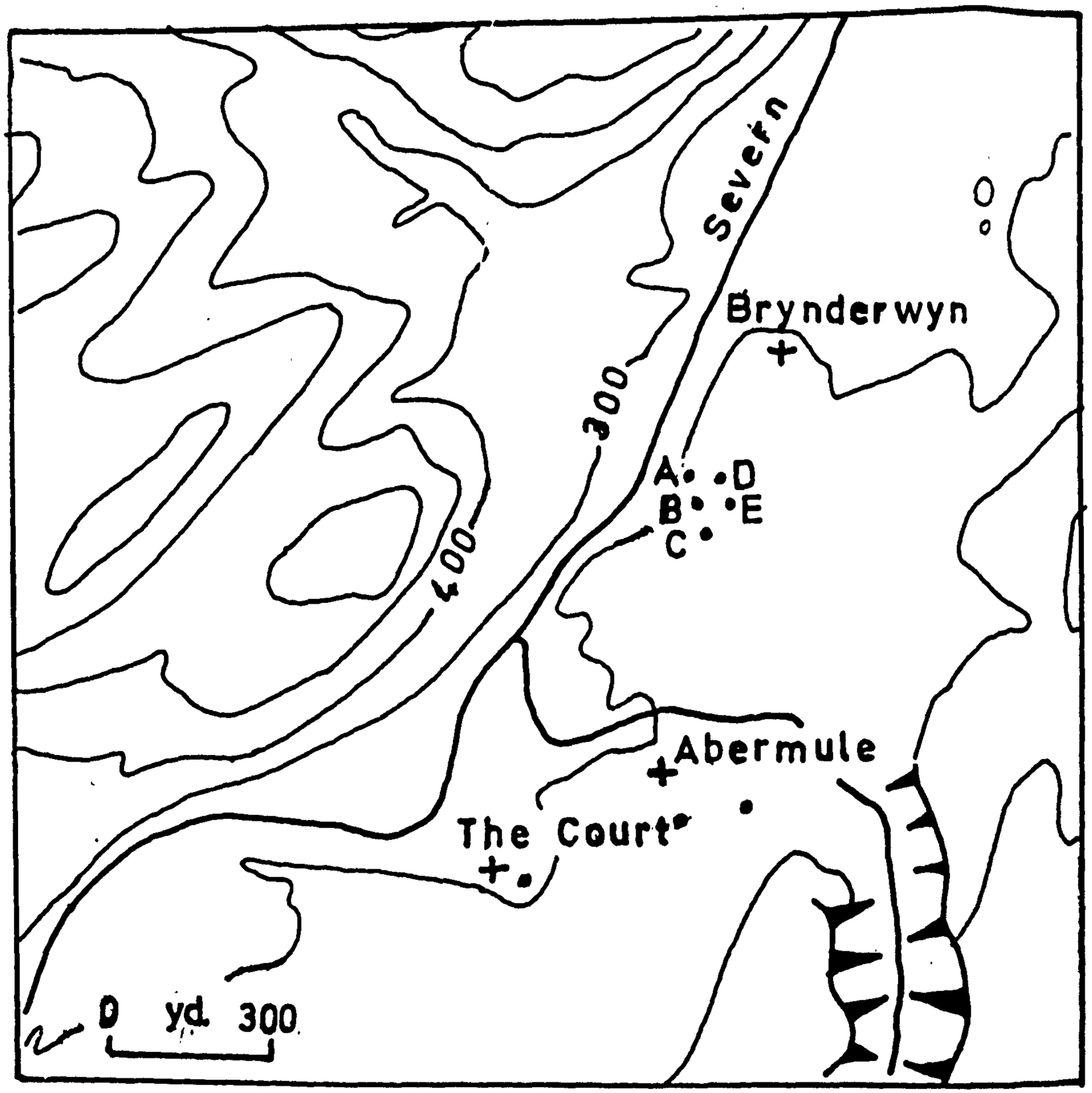


Fig. 3.8 Location of Boreholes at Abermule

grey silty, stoneless alluvium, overlying well-bedded, waterworn gravels (fig. 3.4). Glacial drift is banked up against the south-eastern valley side, and extends beyond Green Lane (SO 174957) into the deep valley which heads past Llandyssil.

At Upper Llegodig (SO 172961) several low elongate mounds extend part-way across the Severn valley; they are aligned parallel to the valley sides and sections show them to be bedrock features with only a thin cover of drift. A short distance downvalley is a similar group of mounds, the most prominent of which extends for three-quarters of a mile from Plas Meredydd (SO 183975). At Bryn-llyn the Severn has cut through a bedrock surface at 300 ft., on which several similar bedrock mounds are developed. The mounds are essentially erosional forms and are interpreted as rock drumlins[1].

At Montgomery Station (SO 205980) there is a section showing a large mass of glacial till, made up of brown clay with included Silurian slates and many rounded boulders of shales, grits and sandstones.

Beyond this point the Severn opens out on its eastern side into the broad Camlad-Marton valley. At the junction of the two valleys the floor is relatively flat and covered by alluvial material overlying bedded gravels, possibly outwash material. On the western side of the valley are two prominent terraces, each 5 ft. high which extend from the mouth of the Rhiw valley; they are composed of bedded, waterworn gravels.

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1. See p. 65

Between Berriew and the Breidden Hills the floor of the Severn valley is relatively featureless and flat; the river meanders over a broad flood-plain covered with alluvium, which overlies bedded gravels (fig. 3.6). There is some glacial drift on the valley sides, but generally these are steep bedrock slopes. That a considerable thickness of drift is present below the floor of the Severn valley just north of the Breiddens is shown by a borehole at Lower Farm (SJ 291160), just north of Criggion, where 24 ft. of marls, sands and gravels overlie 46 ft. of blue clay till.

Middletown Gap

Between the Breiddens and the Long Mountain, the Middletown Gap is floored with deposits of glacial till. Though the thickness of drift is not known within the gap, it is generally only thin, and bedrock outcrops are numerous, especially on the spurs and interfluves. The till comprises a brown clay with included erratics of Silurian shales and grits, together with local volcanic rocks.

Beyond the gap, towards Halfway House (SJ 340116), the area is mantled with drift deposits, and a borehole at Marche Manor Farm (SJ 335105) recorded 20 ft. of boulder clay resting on sand.

## CAMLAD - ONNY VALLEY

Church Stoke to Lydham

The Camlad - Onny valley extends eastwards from Church Stoke towards Plowden at southern end of the Long Mynd, and forms a natural

continuation of the broad Mule - Caebitra valley to the west. The steep edge of the Clun Forest, forms the southern side of the valley, while on the north lies the upland area of Corndon, Stipestones and Shelve (fig. 3.9).

The boulder clay covering the hillslopes on the southern side of the Caebitra valley can be traced eastwards onto the footslopes of the Clun Forest. It forms a continuous cover up to about 600 ft., above which the distribution is patchy up to the watershed at about 1200 ft.

The hillslopes around Pentre (SO 278921) are covered with glacial till. A section in a deep valley to the south of Pentre showed 15 ft. of brown clay with included subangular and subrounded erratics, predominantly of Silurian shales, grits and flags, derived locally or from the west. The upper layers of the till are gravelly, with a large proportion of rounded pebbles set in a gritty matrix. Higher up the valley, to the west of Pentre Cwm (SO 282911), the stream is incised some 10 ft. into a similar deposit.

Farther east, till does not extend much above 700 ft. and the upper steep hillslopes are largely drift-free. However, several erratics of fine-grained igneous rocks were seen on Aston Hill at 1200 ft. A small patch of till is exposed just to the east of the hill, above Whitcot Stile (SO 307896), at 1150 ft.

The hillslopes above Bishop's Castle (SO 323888) are mantled with till to about 900 ft. It is thicker and more extensive in

the valleys which drain eastwards towards the town, and is generally thin on the interfluves.

The glacial drift deposited over the hillslopes to the north of Church Stoke, can likewise be traced eastwards into the upper Camlad valley. The southern slopes of Todleth Hill, above 500 ft. are drift-free. However, a quarry to the southwest of Todleth House (SO 285935) showed 8 ft. of bedded gravels plastered against bedrock. The lower 3 ft. of the section comprised coarse gravels and pebbles, and was overlain by a lens, 9 in. thick, of find gravels and coarse sands. Above this was 4 ft. of gravels and pebbles set in a sandy matrix. The larger fragments are of green vesicular lava and tuff, while sandstones, mudstones, shales, lavas and tuffs occur among the finer material. The igneous erratics are all probably derived from local Ordovian rocks.

East of Todleth House the hillslopes are covered by a variable thickness of till. A section in the banks of a gully south of Llanerch (SO 304938) showed about 10 ft. of a brown clay, with erratics of shales, sandstones and volcanic rocks. Meltwater channels draining southwards near Upper Snead (SO 310930) are also incised 5-15 ft. into till. The drift cover can be traced northwards, up to about 900 ft. near Hyssington (SO 313941), and up the broad valley between Bank and Cefn Gunthley towards the Shelve. A thick deposit of drift occurs at the watershed at Pultheley, between the headwaters of the Camlad and a tributary of the West Onny. A large enclosed depression in the drift was probably formed by the melting of a detached block of stagnant ice

after the main glacier had retreated.

The hillslopes around Roveries Wood are generally drift-free above about 700 ft. but till mantles the lower slopes near Snead (SO 318921). A small exposure showed about 5 ft. of blue-grey till containing abundant locally-derived Wenlock shales with Monograptid fossils, together with some igneous material.

The upper Camlad meanders over a broad, flat flood-plain, which falls from 425 ft. south of Upper Snead to 407 ft. at Church Stoke, a distance of about 3 miles. Numerous sections show a stoneless blue-grey silty alluvial clay, which weathers to an ochreous brown. In the centre of the flood-plain west of Owlbury (SO 314917), is a small outcrop of bedrock at 475 ft.

West of Pentre, on the southern side of the flood-plain, are frequent exposures of well-rounded gravels. Adjacent to Court House Bridge (SO 266923) a section showed 2 ft. of alluvium overlying 3 ft. of well-sorted, horizontally-bedded gravels. Another exposure showed the gravels to be interbedded with stoneless clay. These sections are very similar to those in the banks of the Caebitra below Blue Bell Inn. In addition, there is a notable increase in the numbers of rounded pebbles scattered over the fields westwards from Court House towards the Blue Bell. The gravels are probably outwash deposits laid down by meltwaters issuing from an ice-front as it retreated up the Caebitra valley towards Sarn.

<u>Lydham to Plowden</u>

At Lower Heblands (SO 321909), the flat flood-plain of the

Camlad is replaced by irregular, undulating drift-topography, with many enclosed marshy depressions and some small isolated hillocks. The drift is a brown, stony clay till, with numerous angular shale fragments and a few quartzites and volcanic rocks.

The high ground at 600 ft. around Lydham (SO 337911) marks the watershed in the Camlad - Onny valley. A series of small valleys head eastwards from the Lydham area to become tributary to the upper Camlad. They are all cut into glacial drift. However, there are outcrops of Silurian shale at the southern end of Bromleysmill Pool (SO 332915).

Extending in a large arc from Bishop's Castle, through Lydham Heath, More (SO 343915), and a short way up the Camlad valley, is an area of irregular hummocky morainic topography. The moraine is best developed between Lydham Heath and Bishop's Castle, where complex kames and kettle holes rise some 30 ft. above an undulating drift-covered area. Around More and south of Pitcholds (SO 329930) the kames are more regular and isolated. Sections in the kames show variable drift material; in some there are dirty, unsorted sands and gravels, whereas in others there is a brown clay till with erratics of predominantly local material including shales, quartizites and volcanic rocks. Some idea of the thickness and nature of the deposits in this area is given by a borehole at More; showing 12 ft. of coarse gravel overlying 20 ft. of glacial till.

East of the moraine the valley floor is relatively flat ($1-1\frac{1}{2}^{o}$) for about 700 yd. and sections in the banks of the Onny

show up to 1 ft. of stoneless alluvial material overlying 3-4 ft. of well-sorted, bedded sands and gravels, possibly outwash material. Beyond this is another area of irregular morainic topography, well-developed west of Hardwick (SO 369905) and on the southern side of the Onny valley at Eaton (SO 375895). At Hardwick there are a series of complex kames with numerous kettle holes, some up to 15 ft. deep. Small sections in the kames show dirty, unsorted sands and gravels. At Eaton a group of prominent kames, up to 15 ft. high, rests on the flanks of Oakley Mynd. The thickness of the deposits in this area is indicated by a borehole record at Hardwick, where 62 ft. of gravels and 'stony marls' overlie 42 ft. of soft clay, possibly a till, and bedrock is 104 ft. below the surface, at 446 ft.

The area between Hardwick Hall and the steep western edge of the Long Mynd is covered by an extensive spread of glacial till, extending northwards towards Norbury (SO 365928) and Wentnor (SO 385927), and south towards Plowden (SO 379876). Incised into the till are a number of glacial meltwater channels, separated by rounded and broad flat-topped ridges. In areas where there are no meltwater channels, as for example east of The Barns (SO 384901), is a gently-undulating drift topography with marshy depressions.

Several small sections show the till to be a brown or reddish-brown, weathering grey clay with bands of poorly-sorted or unstratified gravelly clay, and numerous subangular and subrounded erratics. These are predominantly of local shales, but also

include sandstones, quartzites, Pentamerus sandstones, fine grained porphyritic rocks, and several crinoidal limestones. The erratics are generally gravel-sized, but occasionally larger boulders, up to 1 ft. across, were seen.

To the east of Criftin House (SO 380915) a section in the banks of Criftin Brook showed 10 ft. of till deposits. The material was unsorted and became increasingly stony and gritty with depth. Erratics included crinoidal limestone, Pentamerus sandstone, rhyolite, ashes and dolerite, and quartzites, together with local Longmyndian greywackes and grits. A well record at Criftin House showed 66 ft. of glacial drift.

At The Home (SO 377903), situated on a ridge of drift, 5 boreholes showed complex drift deposits. The bore, A, on the summit of the ridge showed 60 ft. of brown stony marls, and blue and grey clay, resting on shale; and that on the flank of the ridge, B, showed 31 ft. of similar deposits. The boreholes adjacent to the river, C - E, indicated clays, gravels, loams and sands to a depth of 24 ft. underlain by a variable thickness of clays and marls. To the south, two boreholes adjacent to the Onny, at Eaton Farm (SO 375895) showed 5 ft. of alluvium and 20 ft. of brown till with thin beds of gravel, overlying Wenlock Shales. Erratics in the till include Longmyndian grits and greywackes, Silurian siltstones, mudstones and limestones, and find-grained porphyritic rocks probably derived from the Shelve area.

The large hollow of Asterton and Wentnor Prolley Moor is covered with brown and reddish-brown till. Scattered over the fields south and west of Medlicott are angular blocks of rhyolite, crinoidal limestone, Pentamerus sandstone and fine-grained igneous rocks. The ridge on which Medlicott stands is devoid of drift, but there are several erratics of rhyolite, and a quartz conglomerate similar to that of the Stiperstones. A borehole west of Robury Ring (SO 3914930Q) revealed 15 ft. of brown boulder clay overlying bedrock (Silurian). Erratics included Longmyndian greywackes and grits, Silurian mudstones and siltstones, quartzites and altered igneous rocks.

The erratics of Pentamerus sandstone were probably derived from the Llandovery outcrop located to the south of the Longmyndian between Wentnor and Norbury, while the rhyolites and other igneous rocks came from the Ordovician volcanics of the Corndon-Shelve district; the quartzite was probably derived from the Stiperstones. It would appear, therefore, that ice streams entered this area from the west and north-west, and that there was a northward movement of ice up the Prolley Moor hollow towards Medlicott.

The extensive spread of boulder clay can be traced onto the southern slopes of Heath Mynd and Norbury Hill, and up the East Onny valley. On the lower slopes of Heath Mynd a broad, gently undulating plateau, the surface of which is between 800 ft. and 1,000 ft., is covered by brown stony clay till. On the slopes of Norbury Hill several platforms between 900 ft. and 1,100 ft. are

also drift-covered. Small sections show a brown till with a varied erratic suite, including local Longmyndian grits and conglomerates, Pentamerus sandstones, Wenlock limestones with brachiopod fossils, volcanic tuffs and rhyolite and quartzites. Above the drift are scattered erratics, including fine-grained porphyritic igneous rocks, some of which measure up to 3 ft. across. On the summit of Norbury Hill, at 1325 ft., are large boulders of dark green volcanic rock with faint striations.

Oakley Mynd

The glacial drift covering the floor of the Onny basin may be traced southwards into the broad gaps on either side of Oakley Mynd. East of Bishop's Castle the flat floor of the Lyndham Heath-Brockton gap is covered with a brown stony clay till, with erratics of shales and a few limestones. On the lower slopes of Oakley Mynd, above The Villa (SO 330877), are two terraces, each 15-20 ft. high, with a steep ($13^{\circ}$) forward edge and a flat ($\frac{1}{2}^{\circ}$-$1^{\circ}$) surface. A deep gully cut through the terraces shows them to be composed of a brown stony till, with erratics of shales, sandstones and a few limestones.

A borehole in the centre of the gap, near Oakley Farm (SO 33318811), indicated over 100 ft. of glacial deposits. About 50 ft. of gravels and sandy clay are underlain by at least 51 ft. of boulder clay; bedrock was not reached and must lie below 460 ft. at this point. On the opposite side of the valley, at The Pines (SO 321877), a borehole showed that 52 ft. of glacial

Fig. 3.9 Drift in the Camlad - Onny valley

drift overlies shales.

To the east of Oakley Mynd the broad gap extending from Plowden towards Lydbury North is covered by alluvium and glacial deposits similar to those described above. At the head of the gap the river Onny is incised about 25 ft. below the valley floor, exposing Wenlock Shales. This suggests that the till in the valley floor is of no great thickness, in direct contrast to the Lydham Heath-Brockton gap.

Whitehead has reported the occurrence of erratics of Stiperstones quartzite and boulders of dolerite, similar to that of Squilver Hill (SO 324933), on Oakley Mynd.[1] This indicates that at some stage of the glaciation there was a south-eastward movement of ice across the Onny valley.

Upper East Onny Valley

North of Wentnor the glacial deposits are generally confined to the valley floor, and the surrounding hills are mantled with 'head' (fig. 3.10). The till is a brown or purple-brown clay with a variety of erratics. At the road junction south of Upper Mill (SO 381941) a section showed 6 ft. of till with erratics of local Longmyndian greywackes and grits, Pentamerus sandstone, fine-grained prophyritic lavas, rhyolite, basalt and some quartzite. The till extends up to about 900 ft. at England Shelve (SO 378954), and the road between Cwnd Cottage (SO 382952) and Bridges (SO 349964) crosses the broad, drift-covered floor of the East Onny valley.

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1. in Greig, D.C. and others, 1968, p. 291.

A section in the river bank south of the Chapel (SO 388961) showed 10 ft. of till with erratics of Longmyndian rocks, quartzites, crinoidal limestone and fine-grained volcanic rocks.

The boulder clay continues westwards up a tributary valley towards Kinnerton Green (SO 369963) to a height of over 1,000 ft. A series of hillocks, formed of sands and gravels, is situated on the northern side of the valley, and at the head of the valley is a spread of similar fluvio-glacial material in which are two small enclosed depressions, probably kettle holes. It is possible that a stream of ice passed eastwards along this valley, moving from the Shelve district towards the East Onny.

North-east of Bridges till may be traced up the Ratlinghope valley to a height of over 1,100 ft. A small section east of Ratlinghope Church (SO 403969) showed 2 ft. of brown clay till, with erratics of local Longmyndian rocks and some quartzites. A roadside section showed 6 ft. of till with similar erratics and some fine-grained igneous rocks.

Glacial till covers the floor of the valley leading northwestwards from Bridges towards Marehay (SO 380983). Between Marehay and the Hollies (SO 378991) it contains a substantial proportion of Stiperstones quartzites, which may have been partly derived by solifluxion from the Stiperstones ridge.

The upper East Onny valley, north of Bridges, is covered with boulder clay. A river bank section south of the School (SO 399978) showed 15 ft. of mottled grey and purplish clay, with erratics of

Scattered deposits also occur on the interfluves particularly in the floors of several cols.

Drift covers the floor and sides of the upper Mule valley. Sections around Cwm Maip show a brown clay with erratics mainly of local origin. Similar sections were seen within the valley as far as Wig. The till cover is generally only thin and there are numerous bedrock exposures, particularly within the deep river gorge. At Ceulannau on the eastern side of the Mule, there is a low mound, some 12 ft. high, composed of fluvio-glacial gravels. A 5 ft. section revealed poorly-sorted sands and gravels which are crudely bedded and separated by occasional layers of brown clay a few inches thick dipping south-eastwards. About 600 yd. downvalley, is another small exposure of sands and gravels. About 1 ft. of fluvio-glacial material is situated at the outlet of a small 'in-out' meltwater channel on the valley side.

Below Wig the Mule valley broadens considerably and the river meanders across a drift-covered floor. The drift extends up to about 800 ft. on the western side of the valley. The steeper eastern side, by contrast, is relatively drift-free, bedrock exposures being frequent.

To the east of the Mule, around Synchant, the upper part of the Miheli valley, is covered with similar glacial till. In the lower part of the valley the drift forms a series of terraces on either side of the river (fig. 3.2). They begin abruptly just above Middle Cwm-y-ddalfa, where they are about 40 ft. above the river, and extend downvalley to about half a mile below Lower Cwm-y-ddalfa. Here, the

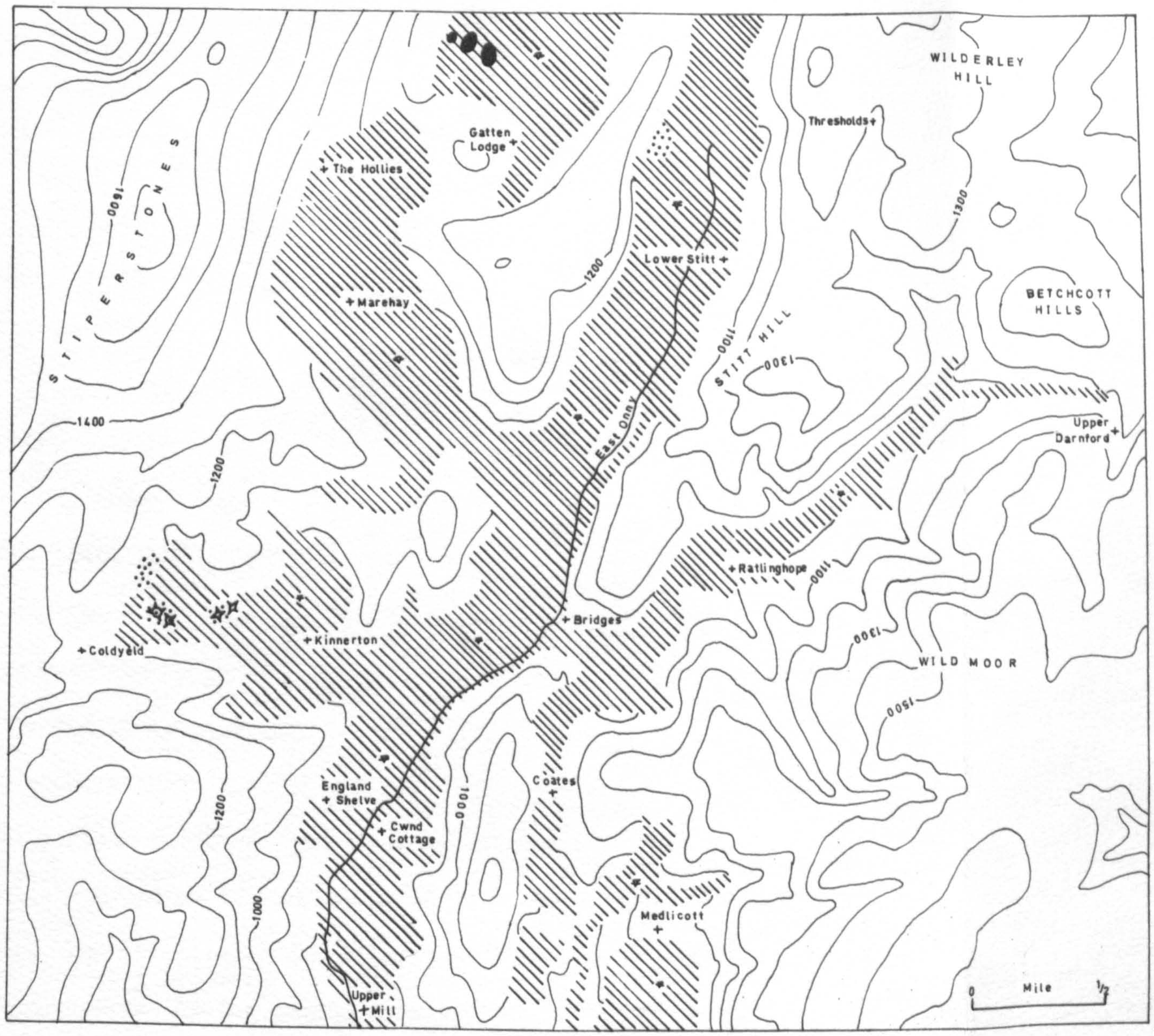


Fig. 3.10 Drift in the upper East Onny valley

Longmyndian rocks, Pentamerus sandstone, fine-grained igneous rocks, tuffs and quartzites. East of Lower Stitt Farm (SO 404986) a section showed a stiff grey and purple clay, weathering yellow, with similar erratics. Southeast of Bank Farm (SO 399997) a small patch of fluvio-glacial sands and gravels has been deposited against the hillside.

According to Cobbold and Whitehead[1] the relatively flat ground between Bridges and Stedment (SO 388969) was probably the side of an ice-dammed lake. There are no exposures of lacustrine clay in this area, and several exposures of till occur within the area that may have been occupied by this lake.

Since erratics of Pentamerus sandstone and crinoidal limestone may be traced continuously up the East Onny valley to beyond Ratlinghope there was probably a movement of ice from the south. The various volcanic rocks and the several quartzites also indicate a trend of ice-movement from the west, from the Shelve area. The erratics in the upper East Onny valley could have been transported by ice coming from either the south or the north.

## CORNDON AND SHELVE DISTRICTS

The glacial deposits covering the area between Todleth Hill and the Roveries on the northern side of the Camlad valley into

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1. in Pocock, R.W., and others, 1938, p. 197.

the valley between Bank and Cefn Gunthly, to a height of at least 850 ft. (fig. 3.11) Above this, the steep slopes of Disgwylfa Hill and Cefn Gunthly are devoid of drift. South of Pultheley (SO 326945) a gully is incised some 30 ft. into drift covering the valley floor, and reveals shales and various local volcanic rocks.

Boulder clay covers the upper part of the West Onny valley, up to about 1050 ft. Again, however, it is confined to the valley floor and the neighbouring hillslopes are drift-free and mantled with head deposits. At Lower Tasker (SO 322964) small streams are incised into the till, and small terraces, 15-20 ft. high, have been cut. Sections show a brown clay till with predominantly local erratics of shales and a variety of igneous rocks, including ashes and tuffs. The terraces continue northwards on either side of the river onto Hyssington Marsh, where they terminate in an area of hummocky drift topography.

Beyond Hyssington Marsh, the drift continues westwards between Corndon and Stapeley Hill. A borehole on the flanks of the former, at North Plantation (SO 305975), showed over 30 ft. of gravels and clay overlying shales.

The boulder clay can also be traced east of Stapely Hill, across Black Marsh, towards the Hope valley. Black Marsh is an extensive drift-covered plateau area at about 1025 ft. Adjacent to Marsh Pool is an area of hummocky drift similar to that on Hyssington Marsh further south. The greater part of the Shelve District is devoid of glacial till, much of the area being covered by head

deposits. However, scattered patches are to be found, usually in valley bottoms, and littered over the hillsides are numerous glacial erratics.

South of Shelve (SO 337990) is a broad, flat-floored valley the floor of which is covered with till, with erratics of shales, dolerite andesite, rhyolitic tuffs and ashes. Adjacent to Shelve Pool (SO 334979) there is an area of hummocky drift with numerous small enclosed depressions. On the southern side of the Pool are two isolated kames each 20 ft. high, in which small sections reveal well-rounded gravels set in a gritty matrix.

Scattered over the hillsides south of Shelve are a wide variety of boulders up to 2 ft. across, including shales, volcanic tuffs, ashes, basalts, andesite, and other medium-grained lavas, rhyolite, pumaceous tuffs and picrite. While the origin of most of the igneous rocks is difficult to determine within this area of complex Ordovician volcanics, picrite outcrops in only a few places, for example at Cefn (SO 312949) in the south-west. This erratic suggests, therefore, a north-easterly movement of ice towards the Shelve district.

East of Shelve, the broad plateau area at the foot of the Stiperstones, known as Pennerley and The Bog, is covered by thick deposits of head. Many sections show up to 4 ft. of yellow-brown clay, with numerous angular boulders, usually 8 in. to 1 ft. across, of Stiperstones quartzite. It seems likely that this material is a downwash or solifluxion deposit from the Stiperstones.

Between Heath Mynd and Linley Hill lies the deep, steep and narrow West Onny valley. The river meanders across a narrow flood-plain, and sections show up to 1 ft. of brown, stoneless alluvial silty clay, underlain by a stony clay with a variety of subrounded gravels. These include shales, sandstones, ashes, tuffs, basalts, dolerite and other fine- and medium-grained igneous rocks. It seems probable that these are stream deposits, rather than of glacial origin since a certain degree of crude bedding was evident. The nature of the material underlying these deposits could not, however, be determined.

Beyond the hills of the Shelve district several valleys drain northwards across the upland area lying to the west of the Stiperstones; the most prominent of these is the glacial meltwater channel known as Hope valley. Much of this area is covered by glacial deposits, which become thicker and more complex towards the north; in the south they are generally only thin, and many of the hills are mantled with head deposits.

The broad depression which heads from Shelfield (SO 343998) towards the Hope valley is covered with glacial till - a brown or yellow-brown clay, with local erratics of shales and various volcanic rocks. Similarly, the depression between Buxton Hill and the Stiperstones is covered by till, which also extends a little way up some of the 'batches'. Generally, however, these are floored with alluvium and head. A borehole at the mouth of Myttonsbeach, just east of the Stiperstones Inn (SJ 364004) showed; that about 44 ft. of head deposits overlie bedrock.

Fig. 3.11 Drift in the Corndon - Shelve district

The upper part of the Hope valley, is incised into bedrock, though the flat valley floor, below about 900 ft, is mantled with till. This is a brown stony clay with erratics of volcanic rocks and shales. The valley to the west is floored by a similar deposit. Below Hope, the valley traverses an area mantled with glacial drift. The higher northern side of the channel is cut through bedrock, but the lower southern side is in drift and slumping of the deposits has occurred in places. At Snailbeach (SJ 373022) are a number of irregular kames up to 25 ft. high. Sections in the flanks of these show unbedded sands and gravels, the latter being generally rounded, but some larger subangular pebbles also occur. The gravels are mainly of local shales and quartzites, together with a variety of fine-grained igneous rocks. To the west, the Hope valley exposes about 20 ft. of fluvio-glacial sands and gravels. At Wagbeach (SJ 365028) is a similar, though less prominent group of kames.

The Hope valley opens out onto a broad drift-covered area extending towards Minsterley. Again, the till is a stony brown clay, with numerous erratics of local shales and quartzites, and igneous rocks. Around Ploxgreen (SJ 367037) several erratics of gfeen dolerite, probably from the Breiddens, to the north-west were observed. It was suggested earlier, that similar erratics from the Breiddens at Asterley and Minsterley indicated a south-easterly movement of ice in this area, and the presence of green dolerites at Ploxgreen would lend support to this view.

## HABBERLEY VALLEY

The floor of the valley, which is bounded on the west by the Stiperstones and on the east by the Long Mynd, is covered by glacial deposits; the till is typically a stiff yellow-brown or brown clay with erratics of both local and 'Welsh' origin (fig.3.12).

The upper part of the valley, separated from the headwaters of the East Onny by an outcrop of bedrock around Gatten Lodge (SO 390992), is covered by glacial till to a height of over 1,100 ft. To the east of Pollardine (SJ 391002), adjacent to the Barytes Mine, a section showed 6 ft. of brown to purple brown clay with many angular and subangular erratics. These are mainly local Longmyndian sandstones and grits, together with some Stiperstones quartzite and Pentamerus sandstone. Numerous specimens of the latter two erratics together with some brecciated and re-cemented quartzites were found scattered over the fields around Pollardine. The presence of Stiperstones quartzites on the eastern side of the Habberley valley indicates a movement of ice across the Stiperstones ridge. The erratics of Pentamerus sandstone could have been derived either from the Upper Llandowery outcrop on the southern margin of the Long Mynd - Shelve district, or from that on its western margin, bordering the Camlad- Rea Brook valley. In view of the presence of Stiperstones quartzites, the latter seems more likely.

Further downvalley, a section in the banks of the Habberley Brook showed 3 ft. of glacial till, a brown clay with erratics of local purple Longmyndian sandstone, grey sandstone, pink and white

quartzite, and Pentamerus sandstone.

At Pollardine is a prominent isolated kame, some 25 ft. high, the surface of which is at 1025 ft. (Pl.10). Small sections in the flank of this feature show dirty fluvio-glacial gravels and sands, with no apparent structure or bedding, and with occasional rounded boulders up to 1 ft. across. Two small patches of gravels also occur on the northern edge of Paulith Bank.

At Marseley (SJ 396023), opposite the mouth of the Eastridge meltwater channel, is a conspicuous mound of fluvio-glacial deposits some 30 ft. high. It has a relatively flat surface at 725 ft., which slopes down towards the centre of the Habberley valley. The northern and western edges of this feature are steep (14°), with sharp breaks of slope, but the other sides are relatively gentle (6-7°) and defined by only changes in slope. On its northeastern edge is a large enclosed depression, 10-15 ft. deep and about 10 ft. across, probably a kettle hole. A 3 ft. trench section in the surface of the mound revealed a very dirty, stony deposit, with sub-rounded gravels and cobbles set in a brown loam matrix. The erratics were predominantly of quartzites, together with some sandstone, including Pentamerus sandstone.

The valley floor towards Habberley (SJ 398036) is covered by glacial till. A section to the west of the village shows 4 ft. of yellow brown clay, included in which are many angular quartzites of cobble size, together with rounded and subrounded gravels of sandstones, grits, and occasional Pentamerus sandstone.

East of Habberley, a section in the banks of the stream near Prestley (SJ 406036) showed 2 ft. of boulder clay, with erratics of local Longmyndian grits and sandstones together with several Stiperstones quartzites, Pentamerus sandstones, and fine-grained basic volcanic rocks, probably basalt. Similar erratics are scattered over the fields to the east. The Pentamerus sandstone and quartzite erratics again suggest a movement of ice from the west or northwest across the Stiperstone ridge. The igneous rocks were possibly derived from the outcrop at Pontesford and Earl's Hills, immediately to the north.

East of Habberley, on the lower slopes of the Stiperstones, immediately to the south of the mouth of the Granham's Moor meltwater channel, is a flat-topped terrace feature some 30 ft. high, with a steep ($13^{o}$) western face. Small sections in the terrace showed it to be composed of fluvio-glacial sands and gravels. North of the village, is a group of irregular kames up to 20 ft. high, in an area of undulating drift topography.

East of Pontesford Hill and Earl's Hills as far as a line between Longden and Longden Common (fig. 3.12), much of the area is covered by a brown or yellow-brown stony clay till, with local and 'Welsh' erratics, similar to that found in the Habberley valley. The nature and distribution of erratics in this area is strongly indicative of a trend of ice-movement from the west or north-west.

A section in the floor of the Radlith col. shows 6 ft. of red-brown loamy clay, with sub-rounded and rounded erratics of local Longmyndian sandstones, occasional quartzites, and several boulders of olivine dolerite probably derived from the Pontesford and Earl's Hills to the west. Similarly, scattered over the fields around Oaks (SJ 423047) are numerous erratics of local purple and brown Longmyndian sandstone, quartzites and dolerites.

Glacial drift of a similar character covers the area to the east of Church Pulverbatch (SJ 430029) and Castle Pulberbatch (SJ 425024), but the hills bordering the Habberley valley are drift-free. Scattered over the fields between Castle Pulberbatch and Lawn Hill are numerous erratics of Longmyndian grits and sandstones, basalts and other fine-grained igneous rocks, and several specimens of brecciated and re-cemented quartzite derived from the Stiperstones to the west. The latter are sometimes of quite large dimensions, and one boulder 3ft. x 1ft. x 1ft. is situated immediately to the north of Castle Pulverbatch. In the village of Church Pulberbatch is a large boulder measuring 4ft. x 3ft. x 2½ft., of green dolerite, very similar to the distinctive green dolerite of the Breidden Hills. The boulder is sub-rounded and smoothed, and bears faint traces of striations.

Southeastwards from Castle Pulverbatch, the glacial drift may be traced up the floor of the broad valley north of Cothercott Hill. East of Cothercott (SJ 420015) is a small mound of fluvio-glacial sands and gravels. Although the valley sides are generally

devoid of drift, small patches do occur, and numerous erratics are scattered over the upper slopes of Cothercott Hill and the adjacent Wilderley Hill, up to a height of 1,350 ft. These include boulders of brecciated and re-cemented quartzite from the Stiperstones and local purple Longmyndian sandstones and grit. An erratic of pink 'Eskdale' granite was found by Cobbold at 1333 ft. at the Thresholds (SO 411994), though Whitehead believes it may have been brought by human agency.[1]

## NORTH SLOPES OF THE LONG MYND

To the east and southeast of Pulverbatch the glacial deposits are significantly different in character from those previously described. While the latter contain erratics which consistently point to a 'local' or 'Welsh' origin, these deposits yield many erratics, particularly granites, derived from a northern source, from the Lake District and Scotland, or from the Trias of the Cheshire-Shropshire Basin. Furthermore, the till is red-brown in colour, presumably derived from the passage of the ice over the Trias. The drifts are the deposits of the 'Irish Sea' ice sheet which pushed its way southwards across the Cheshire-Shropshire basin, onto the margins of the Welsh Borderland.

Adjacent to Underhill Hall and Sheppen Fields is a prominent group of kames, which lie between 800 and 975 ft. (Pl. III). A kettle hole is situated in the flank of a kame to the north of Underhill Hall. A small section close to Sheppen Fields revealed

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1. Pocock, R.W. & others, 1938, p. 195. (footnote)

Pl. 10 Kame deposit at Pollardine

Pl. 11 Kames at Sheppen Fields

unbedded sands and gravels, while a 6 ft. section in the flank of a kame north of Underhill Hall showed coarse, dirty sands and gravels, with a rough horizontal bedding. The gravels included varied local Longmyndian erratics, together with oolitic limestones, volcanic ash, fine-grained volcanic rocks, possibly andesite, and several pink 'Eskdale' granites.

That these kames rest on glacial till is indicated by a deep channel section adjacent to Underhill Hall, which showed about 20 ft. of tough red-brown till clay, with erratics of various sizes, from gravels to boulders up to 1 ft. across. There were many Longmyndian rocks together with Eskdale granites and some quartzites. About 200 yd. southwest of Underhill Hall is a large boulder of purple rhyolite, similar to that outcropping on Pontesford and Earl's Hills. This erratic, together with the quartzites, suggest that the ice from the west mingled its debris with that from the Irish Sea basin. No 'Northern' erratics were found within the drifts of the Habberley valley, but the ice which deposited the latter must occasionally have extended eastwards onto the lower slopes of the Long Mynd.

To the southeast, there is a similar group of kames at Wildreck. At Picklescott a number of ridges and flat-topped terraces have been deposited against the hillside, parallel to the contours. An exposure in the side of one ridge revealed a very gravelly fluvio-glacial deposit, comprising unbedded and poorly-sorted material. The included erratics were predominantly

Longmyndian sandstones, various volcanic rocks, including tuffs and ashes, and occasional Eskdale granites.

Glacial till and scattered fluvio-glacial deposits may be traced northeastwards down the slopes of the Long Mynd. An area of irregular kamiform topography occurs between 500 and 600 ft. around Greenfields. Kames, composed of poorly bedded sands and gravels, rise 25 ft. above a till-covered area. In the flanks of some, as for example north of Coppice House (SJ 445013), are large kettle holes, up to 15 ft. deep, indicating the presence of stagnant ice at the time of formation.

Between Smethcott and Woolstaston is an area of varied and complex drift deposition. Prominent features include well-developed kames, up to 30 ft. high, and other, more complicated fluvio-glacial mounds, and kettle holes situated both in the flanks of the kames and in the surface of the surrounding drift topography. Eastwards, towards Leebotwood the kames are less prominent and the topography more subdued. Sections show the mounds to be composed of red brown sand with included rounded gravels and cobbles. Between the area of fluvio-glacial deposition and kamiform topography are areas of till, and sections show up to 8 in. of dark clay with occasional gravels, underlain by grey stoneless clay.

Erratics in the drifts include local Longmyndian sandstones, flags, grits and conglomerates, Eskdale and Scottish granites, Ennerdale granophyre, dolerites, (probably local or from the west)

rhyolites, (probably Uriconian) and Stiperstones quartzites. This erratic suite is indicative of an ice-movement from the north, from the Irish Sea basin, while the inclusion of Stiperstones quartzites again demonstrates an apparent intermingling with ice from the west.

Some idea of the complexity of drift deposits in this area is given by a series of boreholes put down near Smethcott. One group, across the Betchcott Brook at Walkmills (SO 458993) showed the following succession:

| | | Thickness | Depth | (ft.) |
|---|---|---|---|---|
| 1. | Blue clay and stones | 8-0 | 8-0 | |
| | Yellow clay and stones | 7-6 | 15-6 | |
| | Clay | 29-6 | 45-0 | |
| | Sand | 2-0 | 47-0 | |
| | Rock | | | |
| 2. | (Soil?) | 1-9 | 1-9 | |
| | Gravel | 2-3 | 4-0 | |
| | Sand | 7-0 | 11-0 | |
| | Clay | 23-0 | 34-0 | |
| | Rock | | | |
| 3. | Sandy soil | 4-0 | 4-0 | |
| | Loam | 2-6 | 6-6 | |
| | Gravel | 12-0 | 18-6 | |
| | Sandy clay | 7-6 | 26-0 | |
| | Clay | 12-0 | 28-0 | |
| | Sand | 3-0 | 41-0 | |
| | Purple Clay and Shale below (Longmyndian) | | | |

While another series across the brook south of Branmill (SO 448992) showed:

| | | | | |
|---|---|---|---|---|
| 4. | Clay and stones | 3-0 | 3-0 | |

| | | | |
|---|---|---|---|
| | Gravel | 8-0 | 11-0 |
| | Clay and stones | 16-0 | 27-0 |
| | Gravels | 3-0 | 30-0 |
| | Hard sand | 14-0 | 44-0 |
| | Clay | 12-0 | 56-0 |
| | Blue clay and shale | 20-0 | 76-0 |

| | | | |
|---|---|---|---|
| 5. | Clay | 3-0 | 3-0 |
| | Gravel | 4-0 | 7-0 |
| | Running sand | 14-0 | 21-0 |
| 6. | Clay | 20-0 | 20-0 |
| In part | Loose rock | 2-0 | 22-0 |
| Longmyn- | Shale | 6-0 | 28-0 |
| dian | Clay and Shale | 2-0 | 30-0 |

Whereas, at Smethcott Farm (SO 47359653) a borehole showed:

| | | |
|---|---|---|
| Stony marl | 37-0 | 37-0 |
| Sandy marl | 15-0 | 52-0 |
| Running sand | 1-0 | 53-0 |
| Marl with sand layers | 11-0 | 64-0 |
| Soft marl | 6-0 | 79-0 |
| Soft marl | 3-0 | 73-0 |
| Loamy clay | 13-0 | 86-0 |
| Brown shale and stones | 36-0 | 102-0 |
| Brown shale (loose) | 2-0 | 104-0 |
| Sand layers | 2-0 | 106-0 |
| Brown shale | 1-0 | 107-0 |
| Brown rock | 5-6 | 112-6 |

These borehole records clearly demonstrate the complex alternations of sands, gravels, stony clay and clay which characterise these deposits.

Dorrington - Stapleton Esker

Perhaps the most impressive drift feature in this area is the esker system which trends roughly south-south-eastwards from the

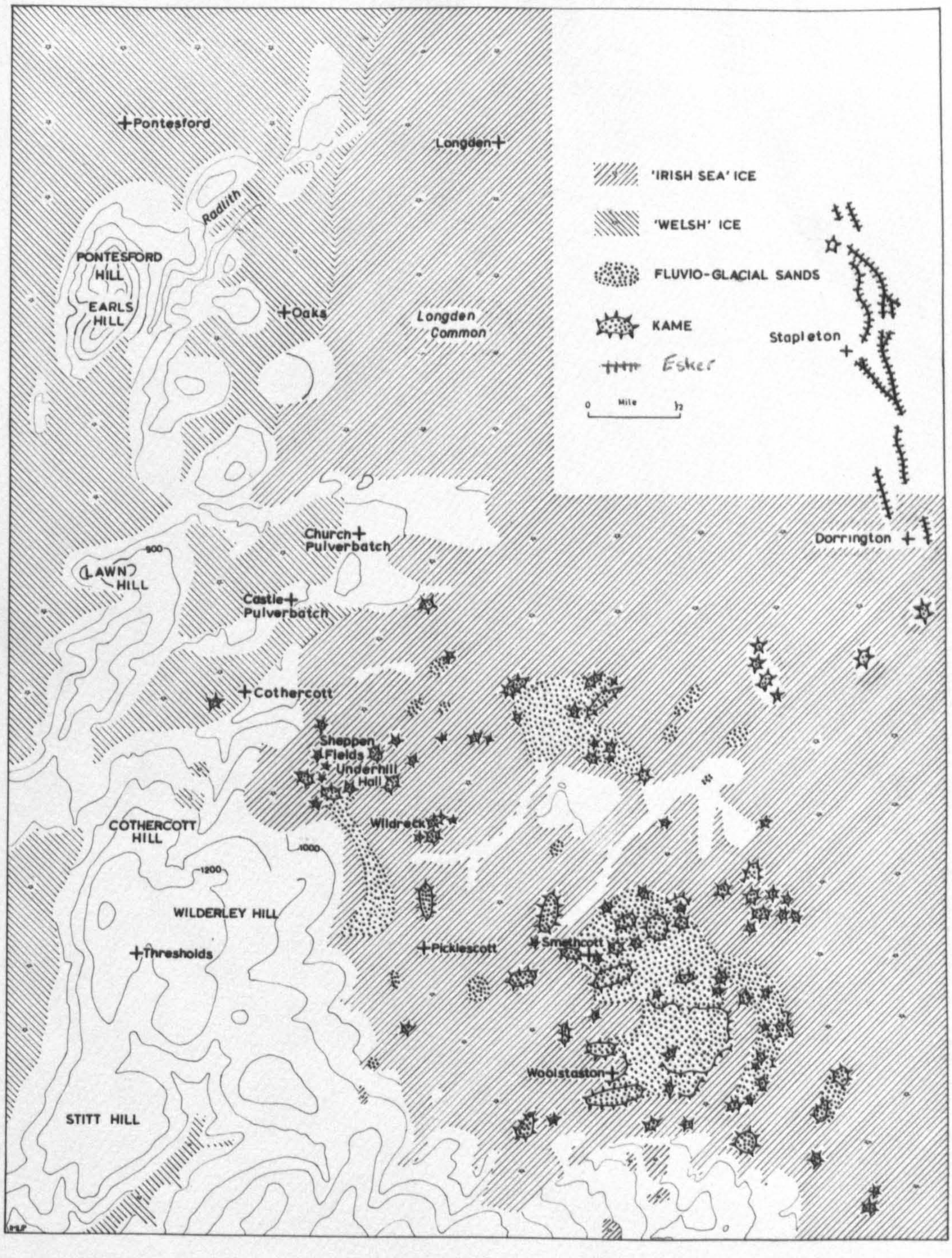


Fig. 3.12 Drift on the northern slopes of the Long Mynd

lower slopes of Lyth Hill (fig. 3.12). Two composite ridges of sand and gravel, each about 2 miles long, are situated on either side of the Cound Brook, extending through Dorrington and Great Ryton respectively. Both eskers are composed of several distinct ridges, rising up to a maximum of 75 ft. above the surrounding plain, which often bifurcate and unite and resemble a stream pattern. That to the west of Cound Brook is best-developed in the north, around Stapleton where two ridges run roughly parallel to each other, and in one place a small kettle hole occurs in the trough between them. Several kettle holes also occur in the surface of the drift between the eskers and Lyth Hill. The other esker is also best-developed towards the north, and the bifurcation of a ridge is well-seen near Condover Grove (SJ 487052). To the south-west of the main esker system a chain of kames, each about 25 ft. high, may be traced south-south-eastwards from Netley Old Hall.

In a large pit adjacent to the road south-east of Stapleton the esker was seen to be composed of bedded sands and gravels, often showing well-developed current bedding (Pl. 12). The dip of the beds was generally down towards the south. Other exposures in the esker system showed similar deposits. In a gravel pit to the south of Dorrington 6 ft. of coarse gravels are underlain by about 4 ft. of fine reddish-brown sand. Pits to the north of Ryton also showed about 6-8 ft. of gravels overlying fine sands. In no case was till seen to overlie the gravels and sands. The erratic content of the gravels was very variable, numerous examples of

northern rocks, as well as local and western pebbles being found. There were several pink 'Eskdale' granites, some up to 1 ft. across, volcanic tuffs, andesite, rhyolite, Longmyndian conglomerates and grits, Carboniferous Limestones, coal fragments, Triassic sandstones, shales and some white quartzite. Several fragments of shells were also observed, indentifiable ones including Turritella and Cyprina.

Some idea of the composition and thickness of the deposits within the eskers is given by several boreholes in the area. At Dorrington, three bores in the esker ridge showed the following sequence:

A. Dorrington Creamery

| | | |
|---|---|---|
| Subsoil | 3-0 | 3-0 |
| Dry gravel | 6-0 | 9-0 |
| Water-bearing gravel | 34-0 | 43-0 |

B. Dorrington (Railway Inn)

| | | |
|---|---|---|
| Start 18-6 below surface | | |
| Clay with sand & gravel | 6-6 | 25-0 |
| Blue clay & small gravel | 23-0 | 48-0 |
| Hard grey rock | 3-0 | 51-0 |

C. Dorrington (Opposite Church)

| | | |
|---|---|---|
| Sandy marl | 25-0 | 25-0 |
| Soft blue shale | 9-0 | 34-0 |
| Clay | 26-0 | 60-0 |
| Blue shale | 18-0 | 78-0 |
| Clay | 12-0 | 90-0 |

At this point it seems that the esker is resting on bedrock. To the north, at Walford House (SJ 476035) a borehole showed that 50 ft. of fluvio-glacial sands overlie 26 ft. of till resting on bedrock.

The relatively flat area around the esker system is covered by up to 6 ft. of loose, loamy, stony till with angular stones up to

6 in. across, many of which are striated. The drift extends up to 400 ft. against the southern slopes of Lyth Hill. The junction of the drift and solid was easily mapped by a well-marked change of slope and vegetation, and in many places by a line of springs. Lyth Hill, and its north-easterly continuation Sharpstone Hill, are drift-free, but numerous erratics are scattered over the surface. These include 'Eskdale' and 'Criffel' granites, dolerites and Triassic sandstones.

## VALE OF STRETTON

The Vale of Stretton is characterised by complex deposits of boulder clays, sands and gravels, which are masked in many places by river alluvium, aggradation fans, solifluxion gravels, and deposits of possible Late Glacial lakes (fig. 3.13). The drift is largely confined to the valley floor, very little occurring above 800 ft. In the north it covers the slopes continuously up to 600 ft. along the edge of the Long Mynd, up to 700 ft. and 800 ft. against the Lawley and Caer Caradoc, and is thickly deposited in the col between the two hills, 22 ft. being recorded in a borehole at Shootrough. Glacial till is absent on the western slopes of the valley between All Stretton (SO 460953) and Little Stretton (SO 443917), but occurs on the east up to 800 ft. against Helmeth Hill, in the col between Hazler and Helmeth Hills, and up to 900 ft. against Ragleth. Above this, the drift is patchily distributed, and there are scattered erratics.

Sub-drift Morphology

Some idea of the thickness and complexity of the drift deposits, and of the form of the bedrock floor is given by a number of borehole and well records (fig. 3.14) and the results of a gravity survey.

In the north, at Gorseybank(SO 471967), a bofehole revealed more than 90 ft. of gravels, sands and marls; while at Botvyle (SO 47359653) the following drift succession was recorded: boulder clay 16 ft.; sand and gravel 14 ft.; boulder clay 8 ft.; solid rocks (Coed-yr-Allt Group). At Gorseybank the bedrock surface must lie below 470 ft., whereas at Botvyle it is at 528 ft.

To the north of Church Stretton a series of boreholes and excavations enable the form of the bedrock floor to be ascertained more clearly.[1] On the eastern side of the valley, at Hillcrest (SO 464945), a well record indicates 20 ft. of boulder clay overlying shale and limestone, and the bedrock surface is at 580 ft. Towards the centre of the valley, in Essex Road, Church Stretton, excavations revealed the solid rising to within 6 ft. of the surface, to a height of 609 ft.; while 30 yd. to the north-west it rises to within 7 ft. of the surface, to 615 ft. In both cases Late Glacial torrent gravels, probably derived from the Cardingmill valley, and organic clays overlie the solid. To the north and north-east of this poiht the rock floor falls away steeply. A borehole at Spring Bank Farm (SO 45719447) showed 86 ft. of glacial sand and gravel alternating with three layers of till, overlain by 6 ft. of head deposits, and in turn overlying solid rock (Hughley Shales), the surface of which

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1. Information regarding the excavations was kindly supplied by the late Mr. T.H. Whitehead.

is at 519 ft. To the south-east a borehole revealed 110 ft. of complex Late Glacial and Pleistocene clays, till and gravels overlying the solid, the surface of which is at 496 ft. Two boulder clays were proved in this borehole; an upper till, 6½ ft. thick, is overlain by 45 ft. of Late Glacial organic clays and gravels composed of Longmyndian material, and is separated from a lower till by 10 ft. of waterlaid gravels. The lower till is underlain by 14 ft. of gravels resting on bedrock. The tills were purple-brown in colour, and include erratics predominantly of local origin, but also a few far-travelled pebbles, such as Eskdale granite, Breidden dolerites, andesite and a few shell fragments.[1]

A borehole to the south of Brockhurst Castle (SO 459939) showed 14 ft. of alluvial clay and Longmyndian torrent gravels, overlying gravels, till, clayey gravels, gravels and sands up to 102 ft. thick, resting on solid, the surface of which is at 482 ft. At Little Stretton (SO 444916) another borehole went through 110 ft. of Late Glacial gravels and clays without reaching solid, which must lie below 480 ft.

Further information regarding the sub-drift configuration of the Vale of Stretton is obtained from geophysical investigations, carried out by the Geophysics Department of the Geological Survey.[2] During 1961 and 1962 gravity readings were made along 25 traverses

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1. Rowlands, P.H., 1966, p. 140.
2. Greig, D.C. and others, 1968, p. 307-321.

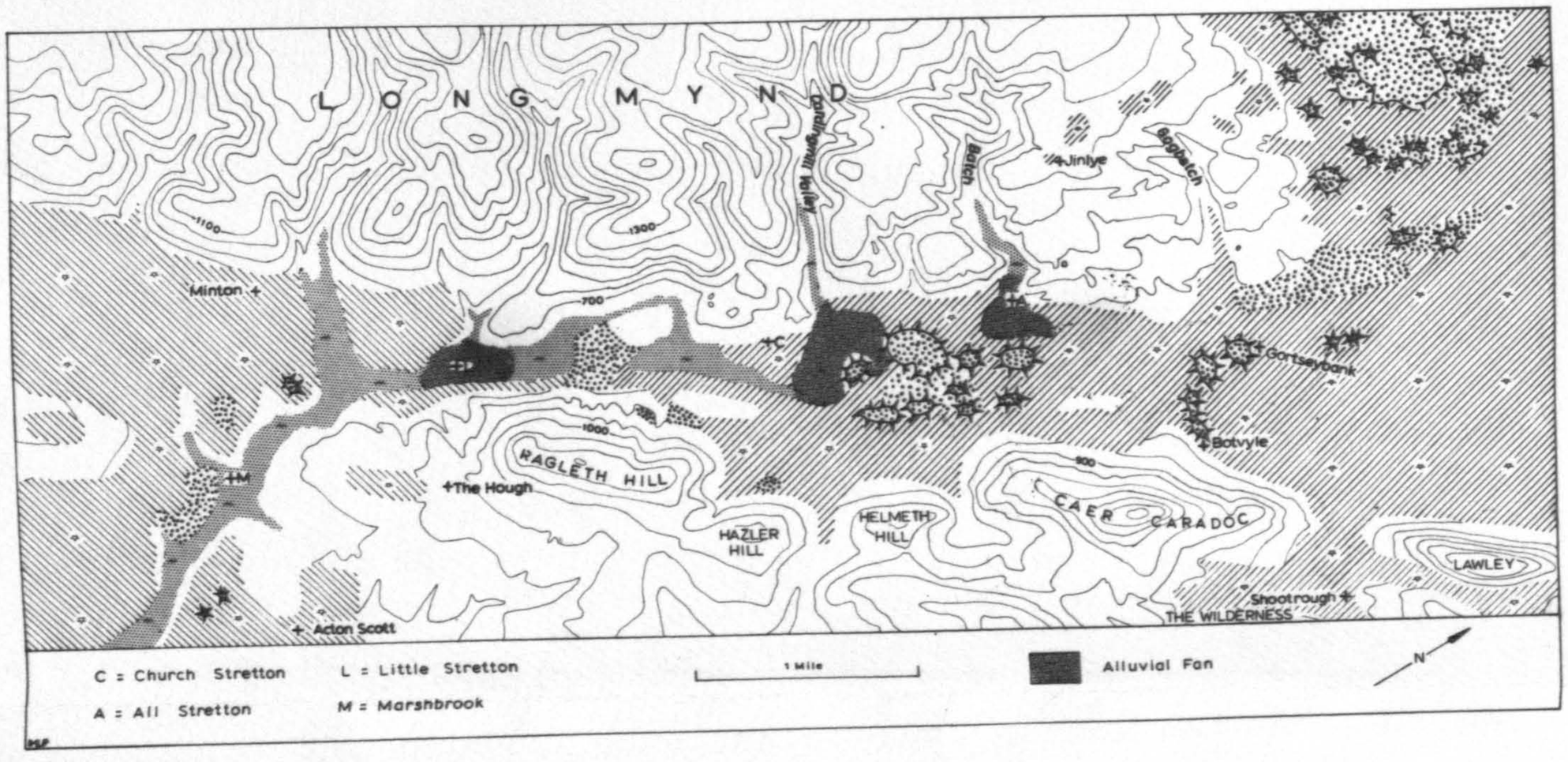


Fig. 3.13 Drift in the Vale of Stretton

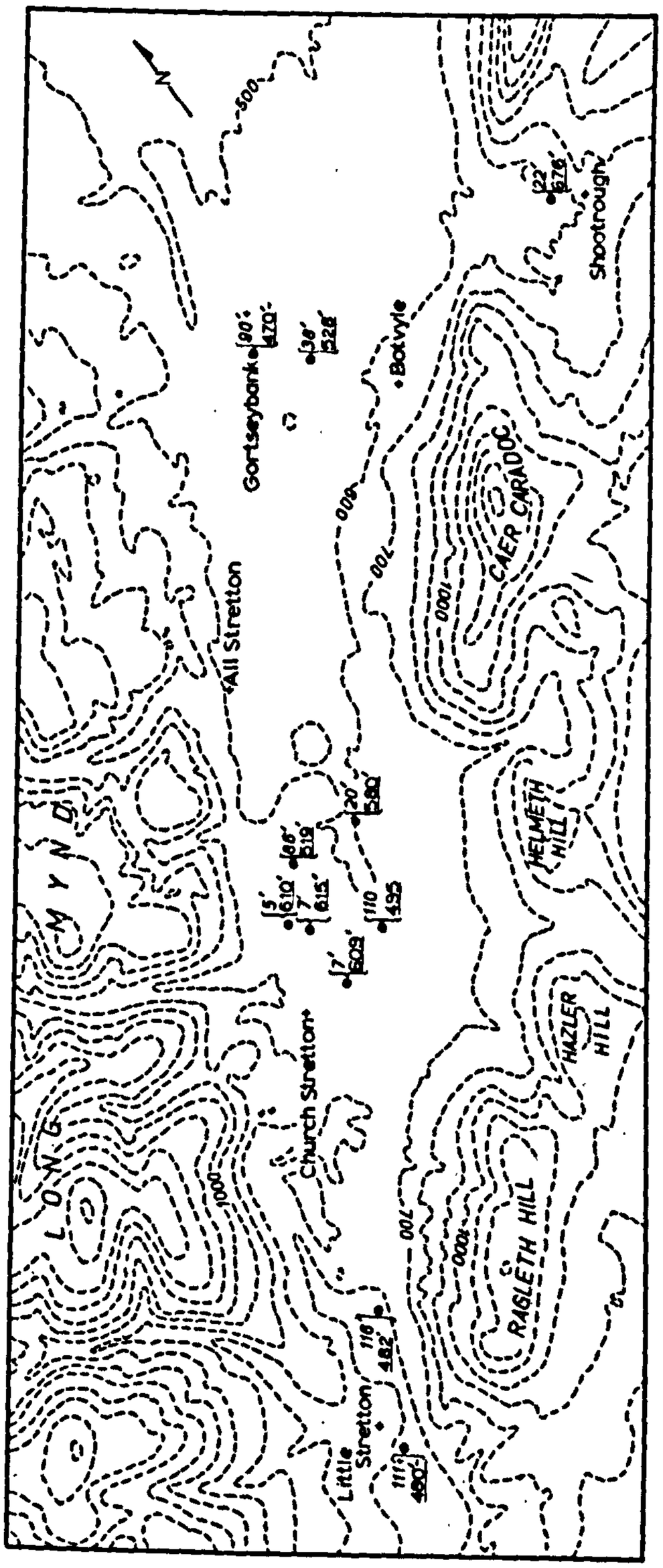


Fig. 3.14 Location of boreholes and wells in the Vale of Stretton

in the Church Stretton valley and at bench marks and spot heights outside the valley (fig. 3.15). On the Bouguer anomaly map the valley stands out as a gravity trough superimposed on a regional gradient of 2 milligals per mile. In places the limit of drift coincides with an inflexion point on the anomaly map, suggesting that the sub-drift surface at the side of the valley is steep, but more often the anomaly lies wholly within the limits of drift.

The local anomaly has its greatest amplitude to the west of Botvyle, and there is a gradual reduction southwards to beyond Little Stretton, where the anomaly diminishes abruptly. Near Marshbrook there is no obvious anomaly. This suggests that there is a southward thinning of drift and alluvium, with the solid near the surface in the Marshbrook gap. The position of greatest negative anomaly follows closely the line of the streams, except for the stream detour west of Brockhurst hill and south of Little Stretton where the stream turns south to pass through the Marshbrook gap.

Valuable checks on the accuracy of the estimation of drift thicknesses by gravity survey are provided by comparing them with borehole records. The Spring Bank Farm borehole falls within traverse 7; 86 ft. of drift were proved in the bore, while the estimate given on the traverse at this point was 90 ft. The deepest part of the valley on this traverse was found to lie c.400 yd. ESE of Spring Bank Farm, at a depth of 150 ft. (fig. 3.15). A well record 200 yd. to the NNE of this point showed only 20 ft. of drift overlying bedrock. The sub-drift surface must, therefore,

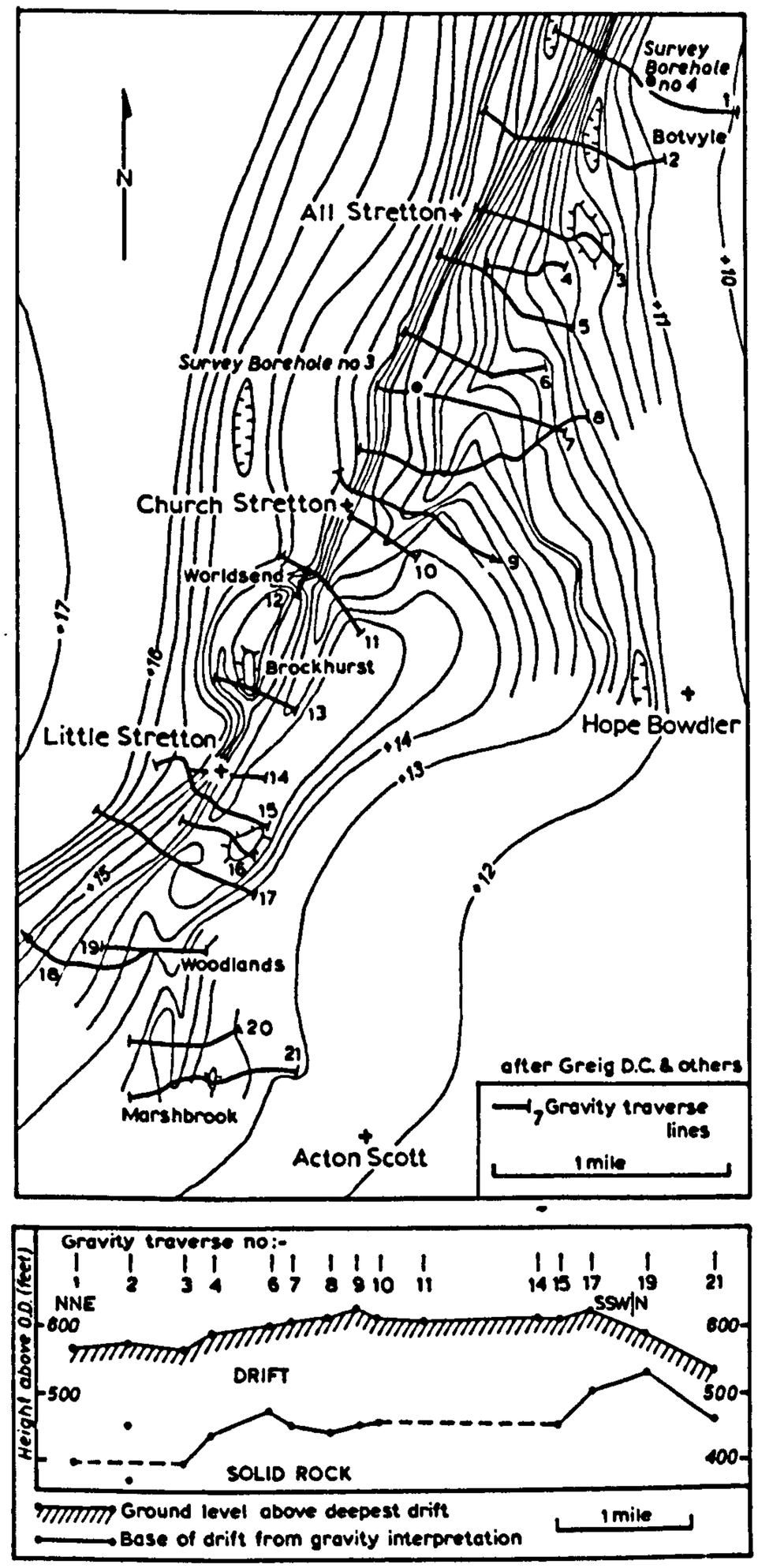


Fig. 3.15 Gravity anomaly map

descend very rapidly to its greatest depth.

The Botvyle borehole falls on the line of traverse 1. The bore proved 38 ft. of drift underlain by about 60 ft. of Upper Carboniferous rocks. The discrepency between the estimate of drift thickness given by the gravity surfey and that proved in the bore is about 18 ft. The greatest depth of drift given by the traverse is 166 ft., close to the western margin of the valley. The bofehole at Gorseybank, 350 yd. NW of Botvyle, proved a maximum of 90 ft. of drift and solid was not reached. This substantiates the thickness of drift estimated on the western side of the valley by the gravity survey.

Traverse 15 indicates a maximum thickness of 155 ft. of drift on the eastern side of the valley near Little Stretton. This is substantiated by the Little Stretton borehole, which penetrated over 110 ft. of gravel without reaching solid. The sub-drift surfaces lies at about 450 ft. at this point. The traverse was taken partly across the mouth of Ashes Hollow and gave an estimate of 25-30 ft. of drift deposits.

At Woodlands, at the entrance to the Marshbrook gorge, the gravity survey (traverse 19) estimated only about 55 ft. of drift and the solid is at about 520 ft. There is thus a sharp rise in the sub-drift profile between Little Stretton and Woodlands.

South of Church Stretton, the present-day drainage west of Brockhurst hill appears from the Bouguer anomaly map to be along a well-established channel. Traverse 12 across the valley at Worldsend indicates approximately 100 ft. of superficial deposits.

This suggests an intimate pre-Glacial or Glacial relationship between the two valleys. The new course may be that of a meltwater channel which eventually cut down nearly as deeply as the main valley east of Brockhurst.

Borehole records and the results of the gravity survey indicate, therefore, that the sub-drift morphology of the Vale of Stretton is complex and irregular. In the north, the deepest part of the valley appears to be on the western side, whereas at Church Stretton and Little Stretton it is deepest on the east. This overdeepening, which is most apparent between Spring Bank Farm and Woodlands, is probably the result of glacier erosion. There is apparently a sub-drift 600 ft. rock platform at Church Stretton, opposite the mouth of the Cardingmill valley. There also appears to be a culmination of the rock floor at 520 ft. at Woodlands, and to 475 ft. to the north of Spring Bank Farm.

Till and Fluvio-glacial deposits ( fig. 3.13)

North of All Stretton, the glacial till is typically a red-brown or brown stony clay to clay loam and, in addition to local erratics, is characterised by pebbles of northern origin, Eskdale granites and Ennerdale granophyre, dolerites, quartzites, and Triassic sandstones. The till is the product of an ice sheet moving towards the south-west, down the Stretton valley. The red colouration and loamy texture is probably largely derived from the Trias of the Shropshire-Cheshire Plain.

Extending across the valley between Gorseybank and Botvyle are a series of gravel mounds, forming a discontinuous ridge. In

the surface of some of the mounds are small kettle holes. Sections show up to 3 ft. of red sand with occasional well-rounded gravels and a few shell fragments. The ridge probably marks a halt stage in the retreat of the Church Stretton glacier. At Botvyle a borehole recorded 38 ft. of drift, and the erratics were mainly of Paleozoic rocks, including Cambrian quartzite and sandstone, grit, Silurian mudstone and siltstone, soft sandstones possibly of Carboniferous origin, and some impure limestones. Erratics of Pre-Cambrian rocks include greywackes and siltstones of Longmyndian type, and altered igneous rocks, possibly Uriconian. There are also fairly abundant examples of dolerite, similar to the Breidden Dolerite.[1]

The hillslopes above All Stretton are largely drift-free and mantled with thin head deposits, but scattered erratics and occasional patches of till at about 1,000 ft. indicate the former presence of ice. There are numerous small pebbles of white quartzite and diorite around Synald's Coppice (SO 457958) and Worsely (SO 455962). At Jinlye (SO 453962) and Bullock's Moor (SO 452970) are small patches of boulder clay. At Plush Hill (SO 451964) is a small patch of till at 1050 ft., around which are scattered numerous erratics, including green dolerite, probably from the Breiddens, white quartzite, quartzite breccia, Eskdale granite, and boulder (2ft. x 3ft. x 2ft.) of fine-grained igneous rock. On the opposite side of the valley

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1. Greig, D.C., and others, 1968, p. 293.

granite erratics were found at over 1000 ft. on the northern end of Caer Caradoc and The Wilderness. These deposits indicate that the ice surmounted the north-eastern shoulder of the Long Mynd, and Caer Caradoc, to a height of about 1100 ft. Ice also pressed eastwards through the gap between The Lawey and Caer Caradoc, depositing drift up to about 800 ft. and the erratics on The Wilderness at 1000 ft.

Close to the streams till is generally overlain by river alluvium. In the Batch valley a stream section showed 3 ft. of alluvial gravels resting on reddish-brown till. The gravels include several erratics, notably Eskdale granites. In Gogbatch a stream section showed 3 ft. of alluvial clay overlying a stiff, stony gravel. The Church Stretton glacier therefore penetrated for short distances up the tributary valleys draining the Long Mynd.

South of All Stretton the valley floor is dominated by a large area of hummocky drift topography, with kame-like mounds rising 30 ft. above the alluvial flood plain (Pl. 13). Adjacent to Spring Bank Farm there are several enclosed depressions within the kames. Sections show red sand with well-rounded gravels, and occasional larger subrounded boulders, with no obvious sorting or bedding. Erratics from the Spring Bank Farm borehole include greywackes, grits and siltstones of Longmyndian type, Silurian mudstones and siltstones, Cambrian quartzites, Uriconian igneous rocks, and Breidden dolerites. In the upper 30 ft. of drift Pre-Cambrian rocks are most abundant, but below about 35 ft. most erratics are

of Paleozoic rocks.

East of Church Stretton, till covers the lower slopes of Hazler and Helmeth Hills, and extends into the col between the two. A section to the south-west of Overdale (SO 457931) showed 5 ft. of stiff brown boulder clay, with erratics of local Longmyndian and Uriconian rocks, together with Eskdale and Criffel granites, granophyre, Triassic sandstone, grits, dark green andesite and white quartzite.

A patch of poorly-bedded sands and gravels occurs at 800 ft. on the lower slopes of Hazler Hill, near Plock's Coppice (SO 463932). Similar deposits are also found on Ragleth, at Cole's Wood (SO 455927), between 700 and 900 ft. Erratics at Coles Wood include grits, quartz and granites.

Thete are few exposures of till on the opposite side of the valley, it being masked by fans of torrent gravels and alluvial clay, but occasional erratics of Eskdale granite have been noted in the lower part of the Cardingmill valley.

South of Church Stretton a ridge of glacial drift is banked up against the southern end of Brockhurst hill. The ridge stretches across the valley in arcuate form, closely resembling an end-moraine; it has flat-topped surface at 700 ft., and a relatively gentle northern side contrasts with a steep ($15^{o}$) downvalley termination. A large gravel pit in the moraine shows about 50 ft. of dirty, poorly-sorted, rudely-bedded gravels with occasional lenses of sand in a reddish-brown, gritty matrix. The gravels are variable

in size, generally ½-5 in. across. The stones seem to be mainly of sub-local origin, though not from the adjacent hills. Two erratics of pink Eskdale granite were found towards the top of the section. In places the gravel was overlain by about 1 ft. of stony clay, probably till. Archeological excavations in the inner ditch of Brockhurst Castle revealed 13 ft. of coarse, poorly-sorted gravel, overlain by 1 ft. of fine gravel, and this by 2 ft. of stony clay, possibly till.[1] The gravel ridge is interpreted as a moraine laid down during a halt phase in the retreat of the Church Stretton glacier. The till at the surface of the ridge may indicate a minor re-advance of the ice.

To the south of Little Stretton glacial drift covers the relatively gentle lower slopes of the Long Mynd, up to about 800 ft., and is locally affected by slumping. On the opposite side of the valley there is a thin cover of drift on the hillslopes between Acton Scott and Ragleth hill. The drift in this area is different in several respects from that to the north. The till is typically a brown or purple-brown clay, and contains a high proportion of local Uriconian and purple Longmyndian erratics, together with others of 'Welsh' origin. Eskdale granites and other erratics of northern origin are absent, particularly towards the south, though several were found in till and fluvio-glacial deposits at The Hough. This area is probably close to the southern limit reached by ice of Irish Sea origin; to the south and southwest the drifts are almost

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1. Whitehead, T.H., personal communication.

exclusively of Welsh origin.

To the south of Woodlands Cottages, a section in a low mound revealed 6 ft. of well-bedded fluvio-glacial sands and gravels, individual lenses of sand up to 3 in. thick being traceable for a distance of up to 5 ft. The erratics are mainly Longmyndian, though some Breidden Dolerites were found. An old gravel pit in a mound west of Acton Scott Church showed a section of 4 ft. of bedded gravels with sand layers. Erratics include purple and grey grits, Acton Scott Limestones, quartzites, tuffs and felsite.

A prominent flat-topped terrace, some 25 ft. high, with a steep ($15^{o}$) face is situated against the western side of the Marshbrook gorge. A section in the northern end of the terrace showed 6 ft. of well-rounded gravels set in a brown sandy matrix; the deposits are poorly sorted but have a rough horizontal bedding. There are a variety of erratics, including Longmyndian purple sandstone and grit, shale, brown sandstones, pink and white quartz, volcanic ash and some Pentamerus sandstone. To the south a sand pit in the terrace revealed 15 ft. of similar fluvio-glacial gravels which include, in addition to the erratics described above, two specimens of pink Eskdale granites.

Discontinuous patches of sand and gravel overlying stony till deposits occur above the western side of the Marshbrook gorge. The erratics include, a few northern granites and several green dolerites, similar to the Breidden Dolerite.

The erratics of Pentamerus sandstone were probably derived

from the Llandovery outcrop around the southern end of the Long Mynd. They suggest a movement of ice from the south-west, up the Horderley-Marshbrook valley. The occasional northern erratics and Breidden Dolerites suggest an intermingling of the drifts of the two ice sheets.

Glacial drift covers the floor of the Marshbrook-Horderley valley (fig. 3.16), extending up to 900 ft. on the relatively gentle lower slopes of the Long Mynd and 800 ft. against the Whittingslow Ridge. The till is a brown or purple-brown stony clay and contains a dominant proportion of Longmyndian rocks, shales, quartzites and Pentamerus sandstones. The colour is derived from the purple Longmyndian and Silurian rocks which outcrop locally. The till cover is generally thin, though a ditch section south of Cwm Head Church (SO 424887) showed 10 ft. of brown stony clay.

Late - and Post-Glacial deposits

Extensive peat and lacustrine deposits occur within the Church Stretton valley (fig. 3.13). Excavations for a new sewerage scheme were undertaken between 1963 and 1964, and they extended from the old sewerage works east of Minton (SO 440910) to All Stretton. The late T.H. Whitehead recorded the sections in the drift revealed by these excavations,[1] and they were studied in detail by Rowlands.[2] At several places, including the vicinity of the old sewerage works, south of Little Stretton, at Worldsend, south-west of Church Stretton

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1. Full details are given in Greig, D.C., and others, 1968, p. 305.
2. Rowlands, P.H., 1966, p. 69-75.

station, and 600 yd north of Church Stretton station, thick deposits of clay and peat were found interbedded with layers of angular Longmyndian gravels. A typical sequence is that recorded at the top of the Church Stretton borehole:

| | Thickness | Depth |
|---|---|---|
| Alternating bands of peat and clay | 11-2 | 11-2 |
| Fine gravel with occasional coarse lenses | 21-4 | 32-6 |
| Grey clay with streaks and bands of black organic clay | 11-6 | 44-0 |
| Grey clay with few pebbles | 0-9 | 44-9 |
| Boulder Clay | | |

The clays and peat are believed to represent a lake, or chain of lakes, which occupied the Church Stretton valley in the Late or Post Glacial period. The spreads of solifluxion material, or flood gravels, are derived from the adjacent valleys of the Long Mynd, such as Callow Hollow and Cardingmill valley, and interrupted lacustrine deposition from time to time.

A radio-carbon date of 11,790 ± 140 yr. B.P. (NPL-81) was obtained on peat material found at a depth of between 5 ft. 8½ in. and 6 ft. in a section to the north of Church Stretton.[1] The stratigraphy of this section is typical of the Late Glacial period. Zone 1 is represented by 2½ ft. of inorganic clays and gravel; Zone 11 by organic clays and p eat, from which the material for C14-dating was obtained; and the overlying clays and gravels are, at least in part, of Zone 111 age. Peats at the top of the

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1. Op. cit., p. 103.

Church Stretton borehole have been dated as commencing at the end of Zone V1 of the Post-Glacial. The grey organic clay underlying the gravel is of Late Glacial age; the gravels were therefore deposited during the interval spanned by Zones 111, 1V, V, and early V1. No dateable horizon within the valley has been found between Zone 11 of the Late Glacial and Zone V1 of the Post-Glacial.

Prominent delta fans, formed of angular stony rubble, occur at the mouths of Ashes Hollow, Cardingmill Valley and Batch valley. At Little Stretton and Church Stretton these fans have spread right across the valley, modifying the northward and southward drainage systems. The accumulation of the delta at Little Stretton probably caused the formation of a shallow lake which almost isolated Brockhurst hill in former times. The fan at All Stretton is less extensive and does not extend right across the main valley.

Thick deposits of angular stony head material occur on relatively gentle slopes on the summit of the Long Mynd Plateau, and has moved little, if at all, since its formation. Excavations and road-side sections show up to 5 ft. of brown silt and stony debris, derived from the underlying rocks. Head is generally absent from the steep sides of the Long Mynd batches, but in the lower part of the valleys there is usually a deposit of angular stony debris which has accumulated partly by solifluxion. The head on the summit of the Long Mynd probably formed under periglacial conditions when the plateau stood as a nunatak above the surrounding ice sheet in the glacial period. Further development and solifluxion of the deposits occurred in a periglacial climate after the retreat

of the ice, in Late Glacial times.

## PLOWDEN - WISTANSTOW GAP

At the southern end of the Long Mynd the river Onny flows eastwards through a prominant gap between Plowden and Wistanstow (SO 432856) (fig. 3.16). Glacial drift of 'Welsh' origin, which is extensively deposited over the Onny basin to the west of the Long Mynd, may be traced into the gap. It occurs on the lower slopes of the Long Mynd, below Park Plantation, and covers the 600 ft. platform.

Exposures to the south of Hillend Farm (SO 398876) and on the 600 ft. surfaces, show the drift to be a stiff yellow-brown or brown clay, with subrounded and subangular erratics of local Longmyndian sandstones and grits, purple shales, mudstones, quartzites and Pentamerus sandstone. To the south of Horderley (SO 408871) a terrace, rises up to 25 ft. above the flat alluvial valley floor. It has a slope of about $3^{\circ}$ towards the centre of the valley, and has a steep ($15^{\circ}$-$17^{\circ}$) forward edge. A section in the terrace showed a stony till-like deposit of angular and sub-rounded stones set in a brown clay matrix.

A similar terrace immediately to the south is composed of well-sorted, bedded fluvio-glacial sands and gravels.

The river is at present meandering over an alluvial flood-

plain, and sections show up to 3 ft. of relatively stoneless silty clay overlying 1'6" of well-rounded and bedded gravels, probably outwash deposits.

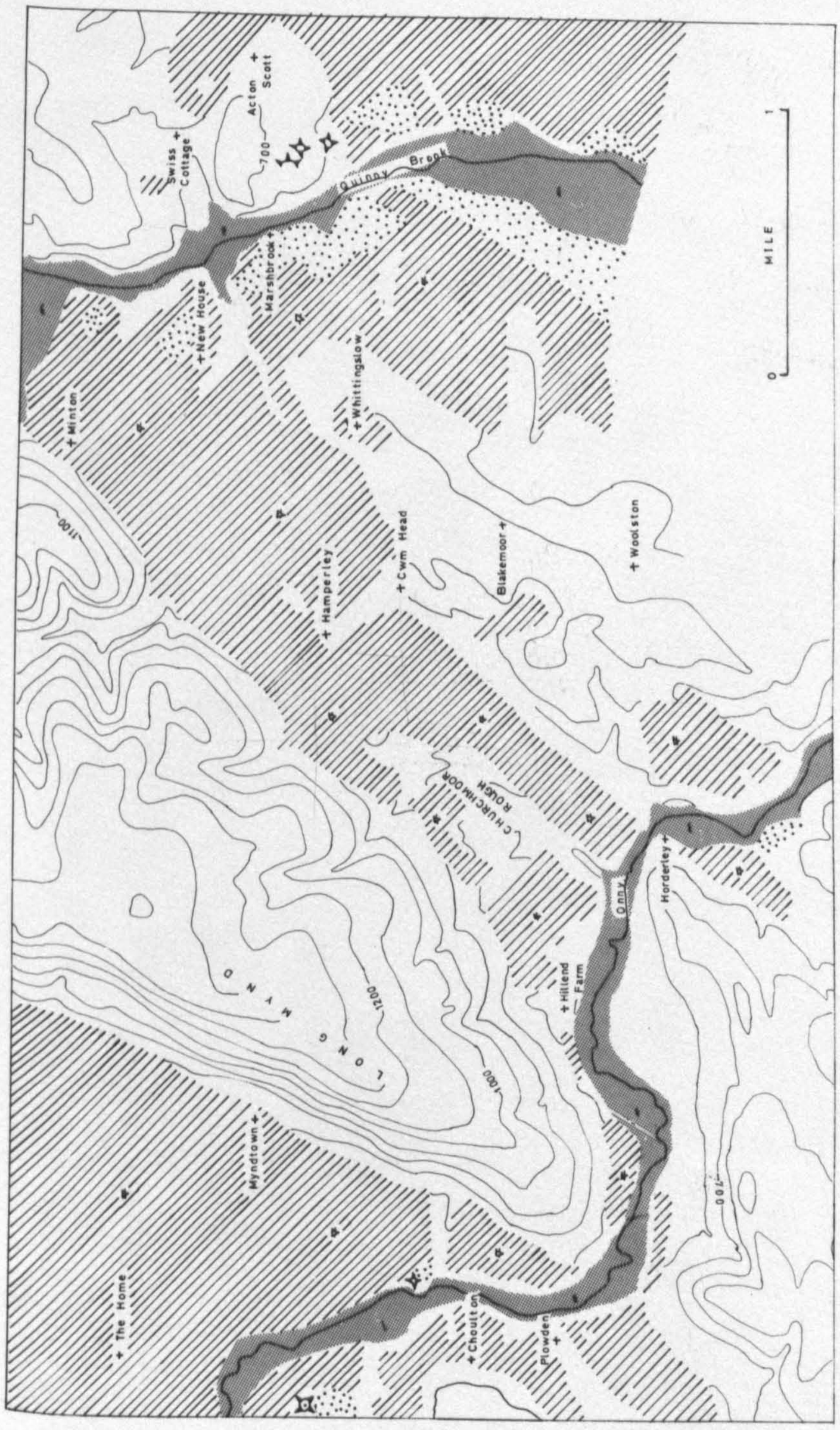


Fig. 3.16 Glacial drift at the southern end of the Long Mynd

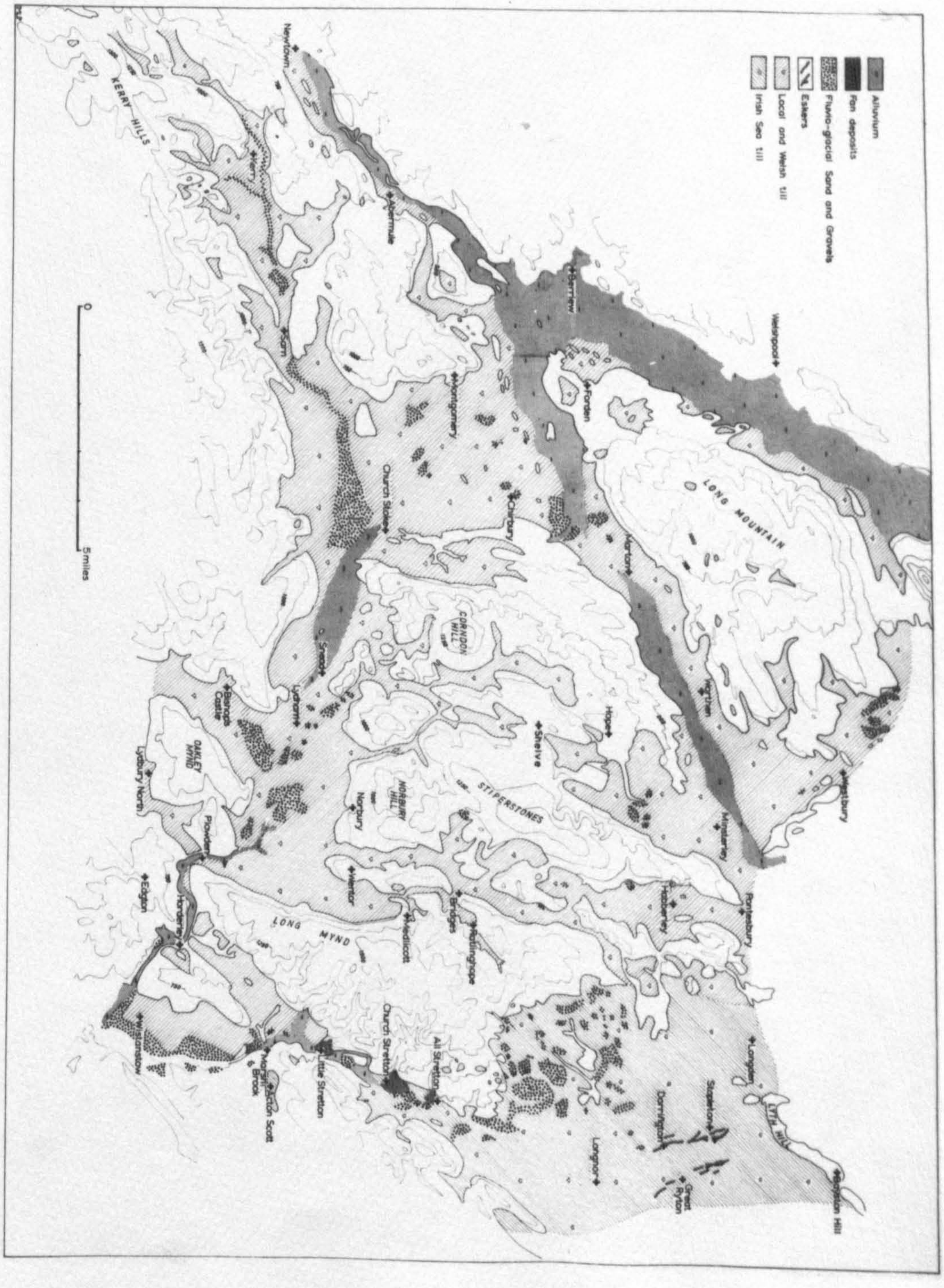


Fig. 3.17 Distribution of glacial drift

# Chapter - 4

## Fluvio-Glacial Erosion

> "Evidence of the former existence of an ice-sheet, as strong in its way as that of striae, is furnished by the presence of rock channels here and there among the hills, excavated by running water in places where the land could have supplied no stream." (Lamplugh, G.W., 1903, p. 363)

Kendall considered glacial meltwater channels to be

> " ... perhaps the most impressive memorials of the Ice-Age that our country contains."[1]

An examination of their form and distribution is essential in any study of glaciated regions since they

> " ... constitute a means by which the behaviour and condition of past ice-sheets can be assessed... (and)... indicated the position and condition of formerly ice-covered terrains."[2]

They provide evidence of the minimum extent of the ice, the mode of ice decay and the slope of the ice surface during deglaciation.

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1. Kendall, P.F., 1902, p. 481.
2. Derbyshire, E., 1962 (a), p. 1111.

## CHARACTER OF CHANNELS

Kendall referred to meltwater channels as anomolous valleys,

> " ... bearing the canyon-like aspect
> that tells of rapid erosion...",

and distinguished two distinct types: where the drainage has been permanently diverted through a channel there is

> " ... a river gorge of a breadth
> disproportionately small as
> compared with the magnitude of
> the stream...",

but where the channels are abandoned,

> " ... the stream, if stream there
> be, is disproportionately small
> for the valley."

Kendall paid little attention to the former type since,

> "Their recognition is easy, and
> their character simple: but the
> deserted channels, having remained
> practically unchanged from the
> time of the departure of the ice,
> with all their characteristics
> unimpaired, are of singular
> interest and demand careful
> consideration."[1]

These anomolous valleys possess a number of peculiarities which distinguishes them from valleys of 'normal' erosion.

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1. Kendall, P.F., op. cit.

"Their transverse section closely resembles that of a railway cutting, with which Prof. Binney aptly compares them..."[1]

"...they invariably exhibit exceedingly steep sides, and, where unmodified by subsequent stream action, possess very broad flat floors:... they indicate rapid cutting by a large stream. The same deduction may be made for the character of the 'meanders'."[2]

"Sinuous channels have a steep batter on the outside bends, and gentler slopes on the inside curve, just as we find in the case of the banks of a river, though not commonly of a river valley."[3]

In long profile,

"The fall...is usually very small near the head and steepens rapidly downstream, a feature which is rarely observed in normal valleys..."[4]

Perhaps the most striking feature is

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1. Kendall, P.F., 1916, p. 27.
2. Kendall, P.F., 1902, p. 484.
3. Kendall, P.F., 1916, p. 27.
4. Kendall, P.F., 1902, p. 483.

> " ...their entire independence of the natural drainage."[1]
>
> " ...they are found cutting through watersheds, even the main water-parting of the whole district..., and sometimes trenching boldly projecting spurs. Often a succession of spurs are cut along a line, as though an originally continuous channel had been segmented by the development of cross-contour valleys of the normal type. A feature which opposes this interpretation is that, instead of the fall-line constituting a continuous gradient from end to end of the series, the outfall of one segment coincides with the intake-level of the next, even though a distance of a mile or two may intervene."[2]

Kendall noted that the 'intake' and 'outfall' of these valleys often display analogous characteristics in that,

> " ...a valley will open out inconsequently on a steep hillside."[3]

It was considered that,

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1. Op. cit.
2. Kendall, P.F., 1916, p. 78.
3. Kendall, P.F., 1902, p. 485.

> "The preservation of the contours of these deserted trenches is, next to their anomolous relations to the relief, their most marked characteristic...whether excavated in granite volcanic rock, slate, sandstone, conglomerate, limestone, shale, or glacial materials, they show in a large proportion of cases scarcely any appreciable signs of weathering; the salient upper edges are only slightly guttered and the re-entrants at the foot of the batter are only rarely filled in by the running down of the banks."[1]

## ORIGIN

Following the publication of Kendall's classic paper on "A system of glacier lakes in the Cleveland Hills", the theory prevalent among British workers for some 60 years was that the Meltwater channels are 'overflows' from ice-dammed lakes. Although Kendall, and other authors,[2] recognised the possibility of meltwater drainage beneath an ice-mass, this was generally regarded as an impenetrable barrier, ponding up glacial lakes and producing overflow channels:[3]

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1. Kendall, P.F., 1916, p. 78.
2. For example: Dwerryhouse, A.R., 1902, p. 585.
3. Kendall, P.F., 1902, p. 480.

> "In all glaciated regions..., the invasion of any district by an ice-sheet, especially where its movement was upstream, would obstruct the natural drainage, producing lakes, the level of which would rise until an outlet for the water was established in some new direction. In some instances the overflow might take place over the advancing ice, or between it and the hillsides in others where the obstruction was of sufficient thickness water might escape laterally, in a direction at right angles to the largest diameter of the lake and to the pre-existing stream."[1]

Mainly using the evidence of 'overflow channels', together with other criteria such as supposed beach deposits, deltas and lake-floor deposits, Kendall identified a series of former glacier lakes in the Cleveland hills, and reconstructed successive ice-front positions during its retreat.

Kendall's views received little criticism and were widely accepted. Numerous, often extensive, glacier lakes were postulated in various parts of Britain, often based solely on the recognition of supposed 'overflow channels'. Also, detailed reconstructions of ice-margin positions were made from the altitudes of supposed lakes and overflow channels. Thus, referring to the glaciation of the northern

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1. Harmer, F.W., 1907, p. 470.

Pennines, Dwerryhouse stated with confidence that,

> "The southern limit (of the ice) ... can be traced with great accuracy by the series of lakes and overflow channels that were produced along that line."[1]

In recent years, however, it has become clear that many difficulties are inherent in the interpretation of meltwater channels solely as overflows from ice-dammed lakes. One of the most profound problems is that complex and often rather improbable positions and fluctuations of the ice-margin are required to explain the position and form of the channels. A further basic criticism is that this interpretation envisages the back-wasting of an ice-front which often pays little attention to local topographic conditions. This is not in accord with conditions existing in present-day areas of glaciation where many authors, including Flint and Desmorest[2], have shown that deglaciation involves the downwasting and stagnating of the ice, rather than the backwasting of an impenetrable barrier. As Mannerfelt, Hoppe, Flint and many others have shown, this

> " ...resulted in a very close relationship between the position of the ice-margin and the form of the ground; the ice remaining longest in the valleys and lowlands. Glacier thinning would have caused the sites of the supposed lakes to have been occupied by ice

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1. Dwerryhouse, A.R., 1902, p. 601.
2. Flint, R.F., 1942.

> at the times the lake should have existed."[1]

In general, British workers understood little of conditions in areas of contemporary glaciation, but as Mannerfelt stated,

> "There is little prospect of understanding and interpreting the often complicated forms (of fluvio-glacial erosion) without a glacialogical view of the late glacial downwasting phenomena that is based on fact."[2]

The 'glacier lake' hypothesis assumes that the majority of channels are formed by overflowing lake waters or by drainage along the edge of the ice; the significance of glacier meltwaters flowing on, in and under the ice is not appreciated. The classic studies of Alaskan glaciers by Russell[3], Tarr[4], and Von Engeln[5], and of glaciers elsewhere by workers such as Garwood[6], and Sharp[7], demonstrated that glacial meltwaters, supplemented by drainage from the surrounding hills, generally flow on, in and under the ice, and erode meltwater channels in a wide variety of positions.

Tarr gave a graphic description of meltwater conditions at the stagnant margin of an ice-mass:

> "Along the margin of every glacier

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1. Sissons, J.B., 1958 (b), p. 182.
2. Mannerfelt, C.M., 1945, p. 223.
3. Russell, I.C., 1893.
4. Tarr, R.S., 1908.
5. Von Engeln, O.D., 1912.
6. Garwood E.J., 1899.
7. Sharp, R.P., 1947.

that reached well down into the zone of melting, in the Yakutat Bay region, there is a marginal channel, with ice for one wall and the mountain side for the other. These marginal channels are noted for their lack of continuity. In some places, after following the margin of the ice for a while, and trimming the mountain slope, they disappear beneath the ice; or they may flow on a gravel bed which rests on buried ice; or they may leave the ice-margin entirely and cut a new gorge across a rock spur, perhaps returning to the ice-margin lower down...Where engaged in gorge cutting these streams work with great rapidity, for the volume is great, the sediment load heavy, and therefore, with sufficient grade to prevent deposit, they are active agents of erosion.

"...above these, marking former higher levels of the ice, are others now abandoned. These are far less easy to understand except in the light of those below, now forming along the existing ice margin. There may be a short section of gorge, contouring the hillslope and open to the air at both ends; or the gorge may extend, on its lower end, across a divide into another valley, and in its lower course appear to be a normal stream gorge though its upper

course is most unnatural without taking into account the former presence of the ice. There may be steep cliffs, evidently trimmed by stream erosion, whose formation is difficult to understand unless one can restore the old ice wall that once formed the other bank of a stream valley. Sometimes these trimmed cliffs end in a gorge like the above. Sometimes they die out without any apparent cause; but this is easily understood when one pictures the stream disappearing under the ice, or flowing on it. At times there is a perfect stream bed at the base of the trimmed cliff; but at other times there is only a narrow terrace, often of very irregular form. Then it requires much imagination to restore a stream valley here where we must supply not only one wall but even the stream bed. The vanished glacier readily accounts for it, however; and if one doubts the explanation, he has but to go a few hundred feet downward to the place where a stream is actually flowing on ice veneered with gravel and walled by a moraine-covered bank on one side and a trimmed cliff on the other. Here it is all clear, for all the elements are there; but in the channel above three of the four elements of the ancient valley are gone - one wall, the stream bed and the stream itself.

"As the ice margin advances and recedes, the position of the marginal drainage

> changes; and when the ice has entirely gone, remnants of the old marginal channels, at different levels, are all that are left of it."[1]

In a similar study, Von Engeln demonstrated that true ice-marginal drainage is an infrequent occurrence, and that meltwater streams usually flow within or under the ice:

> "All valley glaciers probably exhibit some development of marginal drainage features .... Such streams are, however, normally of small volume and short, disappearing under the ice before they can attain any great size....
>
> "Where a valley glacier has existed for a considerable period of time without experiencing any spasmodic disturbance of its mass, whether it is land-ending or tidal, nearly stagnant or in active motion, its lateral drainage will normallly be submarginal."[2]

A number of factors favour submarginal rather than true marginal drainage. Land-ending glaciers typically slope away on both sides from their longitudinal axes, so that at the contact of the lateral edge of the ice-stream and the valley wall, the ice exists as a thin wedge.

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1. Tarr, R.S., 1908, p. 100-102.
2. Von Engeln, O.D., 1912, p. 109-110.

> "Therefore, the comparatively warm drainage from the valley sides can readily melt itself in channel ways under the ice toward the lower portions of the valley; a development which the slow rate of movement, or stagnation, of these glaciers also favours."[1]

Von Engeln also noted that subglacial drainage is normally concentrated at the ice margins and that only rarely does it occur at depth in the central parts of a glacier:

> "...the drainage in all types of the larger glaciers is concentrated in channels submarginal to the ice, while in the smaller, less massive glaciers it may attain a truly central subglacial line of flow. In very large glaciers englacial tubes indicate that a part of the surface drainage may find escape in the central areas of the front by channel ways in the ice, developed, however, at no considerable depths below the surface, perhaps continuing to or existing just below the limiting depth of the zone of crevasses."[2]

Mannerfelt was one of the first workers to apply the observations of people such as Tarr and Von Engeln to formerly-glaciated

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1. Op. cit. p. 111.
2. Op. cit. p. 116.

regions in Europe. In a comprehensive study of depositional and erosional forms associated with deglaciation[1], he recognised four principal types of meltwater channel: Mannerfelt maintained that

> " ...the ice was dead even when the highest peaks in the ice divide area emerged as nunataks."[2]
>
> "Col gullies were formed between the highest peaks. These gullies stand out as dry spillways, often carved out as 10-20 m. deep canyons in the solid mountain. However, they are not as a rule outlet drainage channels for the ice-dammed lakes, but have been formed by running water which flowed along through the mountain passes at the bipartation of the shrinking ice mass. Here, the meltwater ran from the thicker ice through the emerged pass and out towards or above the isolated dead ice on the other side, gradually eroding the bed of the gully "[3] (fig. 4.1).

The water from the col gullies flowed out in some places over the dead ice, and at other places the water was deflected and drained along the ice margin, giving rise to marginal channels. There are also examples of the meltwaters running off under the dead-ice.

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1. Mannerfelt, C.M., 1945.
2. Op. cit. p. 223.
3. Op. cit. p. 224.

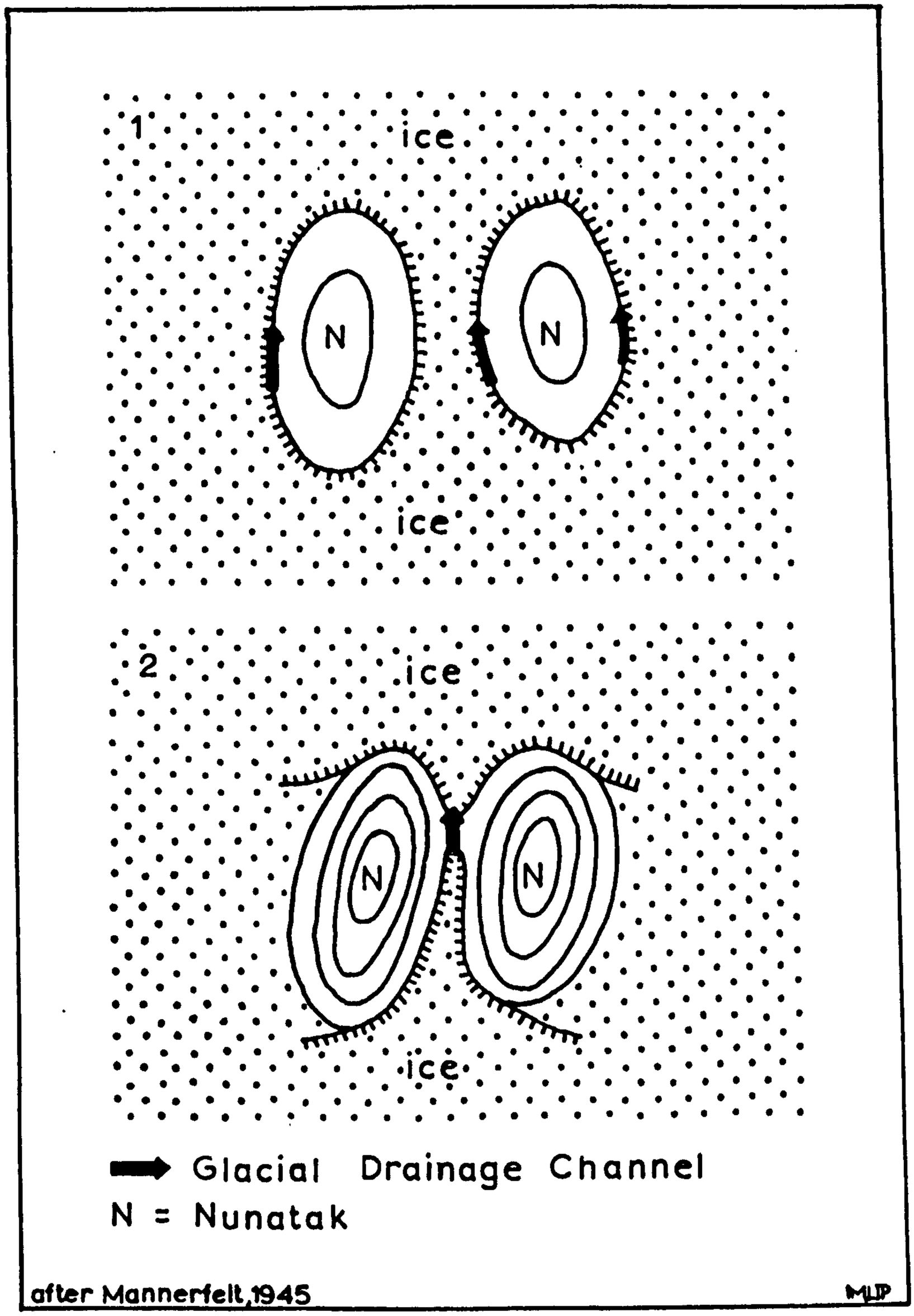


Fig. 4.1 Origin of col gullies, after Mannerfelt

At a later stage of deglaciation, ice tongues remained in the deepest valleys. Thus was formed a whole series of lateral drainage channels along the sides of the valley.

Channels formed between the ice and bared mountain sides are termed 'Marginal drainage channels' but Mannerfelt pointed out that,

> "A typical feature is that the lateral meltwater tries to take a sublateral or subglacial course under the ice margin,"[1]

thus recognising submarginal channels.

Finally, at a late stage of deglaciation when the meltwaters had undermined the dead ice, it was drained off under the ice at the gradient of the local hillside, eroding 'subglacial chutes'.

Following the publication of Mannerfelt's paper, many workers have discussed problems associated with meltwater channels in formerly-glaciated areas in terms of 'stagnant-ice decay', rather than postulating the existence of numerous and large ice-dammed lakes. Foremost among these have been Scandinavians, such as Gilberg,[2] and Holdar;[3] while Sissons[4] has done much to alter the climate of opinion in Britain. Those meltwater channels formed at or near the ice margin are believed to be of particular significance since they afford interesting contributions to our

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1. Op. cit. p. 224.
2. Gilberg, G., 1956.
3. Holdar, C., 1957.
4. Sissons, J.B., 1958 (a), 1958 (b), 1960, 1961, 1963.

understanding of the glacialogical conditions in the final stages of the glacial period.

Marginal channels and benches run along the hillside, roughly parallel to the contour pattern. Benches frequently occur as continuations of well-marked marginal channels, and often, channels and benches alternate one with the other.[1]

Marginal channels and benches may also be recognised by their association with kame terraces and other features for which an ice-marginal origin can be demonstrated.[2]

An ice-marginal origin for these features is partly suggested by their gradient. Mannerfelt is of the opinion that a gradient of 1:100 to 1:50 (1%-2%) is characteristic of marginal features,[3] whereas Gregory gives similar values of 0.75% to 1.5% for marginal features in Eskdale,[4] and Embleton quotes values of 0.5% and 1% for features in North Wales.[5]

They

> "...often occur in sequences, one below the other on hill-sides",[6]

and

> "where gently-inclined channels run approximately parallel with one another for considerable distances a marginal origin seems likely..."[7]

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1. Price, R.J., 1960 (a), p. 83.
2. Gregory, K.J., 1962, p. 369.
3. Mannerfelt, C., 1949, p. 194.
4. Gregory, K.J., 1962, p. 378.
5. Embleton C., 1964 (a), p. 36.
6. Sissons, 1958 (b), p. 161.
7. Sissons, 1960, p. 132.

They enable one to map accurately the former positions of the ice margin, and the gradients give some indication of the superficial slope of the thinning ice mass.[1] As a result of a series of slope measurements in Eskdale, Gregory was able to show that there existed a relationship between the slope of the valley side and the type of meltwater channel eroded:

> " ...it is apparent that ice-marginal benches were cut where the slope was greater than $2\frac{1}{2}^{o}$. Where the angle was less than this, glacial drainage channels developed...On a slope with an angle between $3^{o}$ and $5^{o}$ an ice-marginal bench would be formed first and this might later be developed into an asymmetrical channel."[2] (fig. 4.2).

It is often difficult to prove, however, that any channel was formed strictly at the ice margin. As shown by Von Engeln, simple marginal drainage in present-day temperate glaciers frequently does not exist,[3] and as Price stated,

> "There appears...to be no reason why channels could not have been cut parallel to the ice margin, but subglacially...(and that)... It is therefore difficult to

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1. Mannerfelt C., 1949, p. 194.
2. Gregory, K.J., 1962, p. 370.
3. Von Engeln, O.D., 1912.

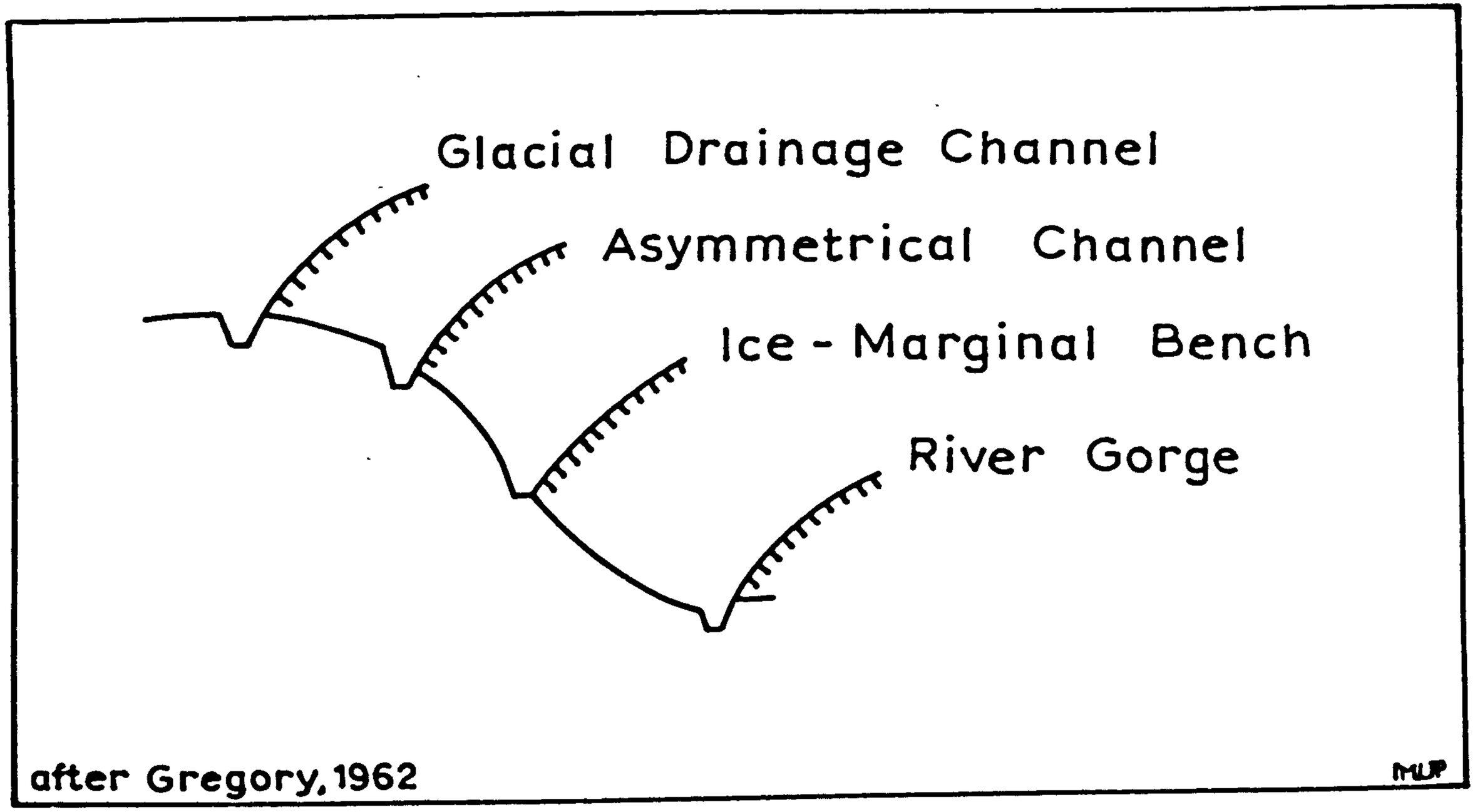


Fig. 4.2 Relationship between channel type and position on a slope

> prove that any channel was formed at the lateral ice-margin."[1]

Gilberg is of the opinion that,

> "Most marginal forms are sub-lateral",[2]

and Bowen has concluded that,

> "...no ice-marginal channels occur in central South Wales."[3]

While the latter may be a rather extreme view, it is clear that many channels believed to be of marginal origin may, in fact, have been eroded submarginally.

Submarginal drainage channels are probably more common and they are generally deeper and more marked than marginal features. They are formed by meltwaters which

> "...found an oblique way down underneath the ice margin,"

and thus have a steeper gradient than marginal channels and benches.[4] Both Mannerfelt and Gregory considered that submarginal channels have a slope of greater than 3.5%,[5] while Embleton quotes a gradient of about 2%;[6] these figures

> "...hardly represent the superficial gradient of the ice, but rather the steeper

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1. Price, R.J., 1963 (a), p. 133.
2. Gilberg, 1956, p. 446.
3. Bowen, D.Q., 1965, p. 347.
4. Mannerfelt, C. 1949, p. 194.
5. Ibid.; Gregory, K.J., 1962, p. 378.
6. Embleton, C., 1964 (a), p. 36.

gradient of local erosion."[1]

A submarginal origin is also sometimes indicated by the sudden changes in direction of many channels, which Sissons believes reflect control of the submarginal drainage in part by structures in the ice.[2] Submarginal channels often form complex anastomosing systems on gently-sloping hillslopes.

Subglacial chutes often originate from marginal and sub-marginal channels, when meltwaters suddenly turn round to plunge downhill beneath the ice in a serpentine course.[3] They generally head directly down the hillsides, more or less at right angles to the marginal channels, and Embleton quotes gradients of 4% - 12.15% as being typical.[4] Sissons states that,

> "...in addition of occurring singly, subglacial chutes frequently occur in groups, sometimes as part of quite complex channel systems, ...(and)...In some localities chutes occur in large numbers with only a limited development of marginal and sub-marginal drainage."

It is not denied that ice-dammed lakes existed during the Pleistocene, or that meltwater channels were eroded when they were drained. As Sissons stated,

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1. Mannerfelt, C. 1949, p. 194.
2. Sissons, J.B., 1960, p. 131.
3. Mannerfelt C., 1949, p. 194.
4. Embleton, C., 1964, p. 36.
5. Sissons, J.B., 1960, p. 133.

> "...it must be made clear that it is not suggested that no ice-dammed lakes ever existed in upland Britain."[1]

> "In certain parts of the world there exist quite large lakes which are dammed up by glaciers, and it is certain that during the various stages of the Pleistocene epoch many others have occurred."[2]

Many ice-dammed lakes exist at the margins of contemporary ice-sheets in Iceland and British Columbia, but these are generally drained off by streams flowing beneath the ice, rather than sub-aerially. The lakes empty catastrophically from time to time, and Thorarinssen

> "...supposes that the water rises until it is nine-tenths of the height of the ice-barrier, after which it can raise the barrier and escape underneath."[3]

Glen, however, has shown that -

> "...an empty hole in a glacier will fill if it is more than some 15-20 m. deep due to the rapid variation in strain rate with stress. It is pointed

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1. Sissons, J.B., 1958 (b), p. 185.
2. Glen, J.W., 1954, p. 317.
3. Op. cit. p. 318.

> out that a water-filled hole
> or lake will enlarge itself
> by the same mechanism, if it
> is more than 150-200 m. deep.
> This phenomena is believed to
> be responsible for the periodical
> and complete emptying of some
> glacier lakes..."[1]

He considers Thorarinssen's theory to be unlikely, and favours the view that

> "...the lake will dig itself a
> tunnel under the glacier until
> either it links up with a sub-
> glacial stream already present
> or emerges at the snout."[2]

Gilberg, writing on the deglaciation of part of the Swedish Highlands, found evidence for the former existence of numerous marginal lakes along the mountain slopes during the downwasting of the ice-sheet.

> "Even when the highest points emerged
> as nunataks from the ice cap, the
> meltwater gathered between the ice-
> free land and the ice, forming small
> lakes. The later development seems
> clearly to indicate a deglaciation
> by successive thinning. Larger and
> lafger land areas became ice-free
> and the ice was increasingly
> concentrated in the valleys. The

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1. Ibid
2. Op. cit. p. 317.

primary nunatak ice-lakes changed during this process into numbers of 'unilateral' marginal ice lakes which were dammed up by the ice itself. When the latter was thin and in the process of disintegration the meltwater usually stood at the same level along the sides of the valleys and 'double' marginal lakes were formed. During this period of shrinking, dead ice was often left behind in small valleys."[1]

He maintained that, "The outlets of most glacial lakes have been subglacial", forming submarginal channels or chutes, but in many cases they were discharged englacially.[2]

Thus, while the existence of ice-dammed lakes during the Pleistocene is not ruled out, it seems that they would have emptied subglacially or englacially, rather than subaerially as envisaged by Kendall, and it is likely that they often occurred in association with downwasting and stagnating ice masses.

## CHANNEL MORPHOLOGY

A consequence of the new and increased study of meltwater channels has shown that they possess a large variety of form, often demonstrating complex origins. In a detailed study of transverse profiles, Gregory has shown that meltwater channels do not always

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1. Gilberg, G., 1956, p. 450-451.
2. Op. cit. p. 447.

typically posses steep sides and broad flat floors:

> "In some the floor is flat and wide, but in others it is V-shaped; these two types have been described in different parts of the Beldon channel, Northumberland... In Eskdale there is a third type in which a flat channel floor has been dissected by a smaller, V-shaped channel."[1]

Gregory also noted that some channels posses narrow benches cut into their sides, thought to be remnants of former channel floors. Ives noted a similar phenomena in that some have a 'channel-in-channel' form, where a more recent channel has been cut into the side of an older one.[2]

It has also become evident that Kendall's statement that the long profiles of glacial meltwater channels typically begin with a low gradient and steepen rapidly downstream,

> "...is not true frequently enough to become a diagnostic feature."[3]

> "This broadly convex (immature) profile is often directly attributable to structure, such that the detailed pattern is one of risings and fallings, sometimes associated with potholing. Theoretically, there is no reason why the long profiles of some

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1. Gregory, J.K., 1962, p. 365.
2. Ives, J.D., 1960 (b), p. 555.
3. Derbyshire, E., 1962,(a), p. 1117.

> drainage channels should not be concave, for example, in a deep, well-established, pre-glacial valley used by a sub-glacial torrent....Many channels possess interrupted profiles, all of which can be ascribed to rapid changes in the resilience of the underlying bedrock."[1]

A feature peculiar to meltwater channels, which has received a great deal of attention in recent years, is the widespread occurrence of an 'up-and-down' or 'humped' long profile. In some cases this is attributable to an infilling of peat, an accumulation of slope wash from the sides, or some other depositional feature; but in many cases the forms are of original erosional profiles. Such anamolous profiles were recognised by Kendall in two channels in the Cleveland Hills, and were explained by the

> "...alternate flow of lake waters ... through the same channel, due to fluctuations in the ice-front giving outlets at both ends of the escarpment."[2]

Peel, also, originally explained the 'up-and-down' profiles of two channels in Northumberland, Beldon Cleugh and East Bipton, by

> "...a reversal in direction of the lake-overflow drainage through the channel occasioned by the drainage away from the original lake and the

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1. Op. cit., p. 1117-1118.
2. Kendall, P.F., 1903, p. 40.

> creation of a new lake on the former outfall side by changes in ice-front positions."[1]

However, as Peel stated, this explanation

> "...requires special conditions unlikely often to be fulfilled."[2]

and although reversal of flow may account for the 'up-and-down' profiles of some channels, there are many others where it appears to be an unsatisfactory explanation.

Among the theories put forward, Twidale proposed that the profile might be formed by the erosion of a channel across a col previously lowered by a pre-glacial stream working backwards by headward erosion, resulting in the floor of the 'pre-glacial' valley forming the uphill part of the profile.[3] This,however, is likely to occur only under rather special circumstances.

None of the above theories provides a satisfactory explanation for 'up-and-down long profiles,

> "...but these difficulties may disappear when the importance of drainage into the ice is recognised."[4]

Sissons proposed a further alternative, that such profiles are produced

> "...by subglacial erosion by water flowing under hydrostatic pressure."[5]

It is only since the publication of Sisson's paper, which is based largely on the work of Mannerfelt, that this hypothesis has received

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1. Peel, R.F., 1956, p. 484.
2. Ibid.
3. Twidale, C.R., 1956 (a).
4. Sissons, J.B., 1960, p. 145.
5. Ibid.

serious attention in Britain. Peel later accepted this origin for the profiles of the two Northumbrian channels, but stated that,

> "Many problems are...involved in such a concept including the ability of subglacial streams to erode large channels in solid rock, (and) the possibility of their eroding uphill under hydrostatic head ..."[1]

Sissons admitted that,

> "Proof that subglacial streams are eroding their beds, especially uphill under hydrostatic pressure is difficult to obtain ...."[2]

but cites examples noted by Okko, Arnborg and Thorarinsson in Iceland that contemporary

> "...subglacial rivers have a great capacity and eroding force."[3]

Although this appears an attractive explanation of many 'up-and down' long profiles in that subglacial streams do flow uphill under hydrostative pressure, it is not clear how the anomalous profile actually originated. Price perceived a connection between the superimposition of meltwaters across a ridge and the formation of an 'up-and-down' long profile:

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1. Peel, R.F., 1956, p. 486.
2. Sissons, J.B., 1958 (c), p. 174.
3. Ibid.

> "A possible sequence of events leading to the development of an 'up-and-down' channel could begin with the superimposition of a large englacial stream on to the crest of a ridge or spur aligned at right angles to the meltwater stream. If the meltwater stream cuts its channel vertically downwards in the ice faster than in the rock forming the spur, the normal superimposed channel with its constant width and gentle gradient would be superceded by a channel with an up-down profile. This hypothesis does not suggest that all the erosion necessary to cut the large channels with up-down profiles was carried out by stream flowing under hydrostatic pressure."[1]

Under this hypothesis, all the erosion required to cut large channels with'up-and-down' profiles was not necessarily carried out by streams flowing under hydrostatic pressure. However, as Price makes clear,

> "The meltwater stream then had to flow uphill under hydrostatic pressure if it was to continue along its course through the spur."[2]

As there are many variations in the cross - and long profiles of meltwater channels, so there are many types of channel inlets

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1. Price, R.J., 1960 (a), p. 128.
2. Op. cit. p. 129.

and outlets;

> "...depending on the bedrock as well as the position and condition of the ice, the intakes of glacier drainage channels vary considerably...."[1]

Derbyshire recognised four principal types of intake. The 'open trumpetlike' type are said to predominate, many of which 'hang' on the valley sides; the precipitous, or 'amphitheatre' type

> "...begins suddenly, plunging down to the main channel level, often in an armchair hollow;"

the 'bench' type represent an incipient stage in channeling; and the 'round stream' type of intake begins suddenly and increases rapidly in breadth downstream.[2]

Derbyshire also recognised two principal types of channel outlet; 'trumpet-shaped' which are the commonest type; and 'benchlike' outlets, which usually represent an incipient stage in the development of a marginal or submarginal channel.[3]

The role of structure in determining the morphology of meltwater channels has often been underestimated, but,

> "The bedrock structures (often) closely control the topography and can be expected to exert strong lithological influence on the morphology of water-cut glacial features."[4]

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1. Derbyshire, E., 1962 (a), p. 1115.
2. Derbyshire, E., 1962 (a), p. 1115-1117.
3. Op. cit. p. 1117.
4. Ives, J.B., & R.P., Kirby, 1964, p. 917.

The nature of the pre-glacial landscape must also be considered since,

> "...it is often not clear whether the features are glacially modified preglacial features or entirely glacial and, in either case, if they have been modified by glacial meltwater during various glacial phases of the Pleistocene."[1]

From the above discussion it is clear that glacial meltwater channels may occur in a wide variety of positions and possess a very varied morphology.

> "The channel position relative to the melting ice mass is determined by climate and glacial conditions - the thickness and temperature of the ice and its regime. The final morphology, however, is decided by these factors in dombination with local topography, lithology and structure. The physical form of any glacial drainage channel ... may be due to many different factors, but the relative importance of each of these factors is not always determinable."[2]

It may be concluded that

> "In view of the lack of precise work on the mechanics of channel

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1. Op. cit. p. 917-918.
2. Ives, J.D., & R.P., Kirby, 1964, p. 917.

> formation and ignorance of the relative importance of the variables involved, it seems unwarranted to modify, enlarge, or further divide Mannerfelt's simple classification,"[1]

and it may be agreed with Sissons that,

> "...owing to our present inadequate knowledge, ...to attempt a classification of channels, or to discuss all types of channels...is not advisable or practicable."[2]

## AGE OF CHANNELS

It is often tacitly assumed that all meltwater channels in a particular area were eroded during the same glacial episode and little account is taken of the effects of earlier glaciations. It is generally agreed that several glacial periods affected the Northern Hemisphere during the Pleistocene, and it seems likely that earlier glaciations must have left some relics of fluvio-glacial erosion.

There is often considerable difficulty in attempting to assign

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1. Op. cit. p. 920.
2. Sissons, J.B., 1961 (a), p. 15.

meltwater channels to a particular glacial period. Kendall was of the opinion that a variation in the "freshness" of the morphology of meltwater channels was related to their origin in different glaciations:

> "The distribution and state of preservation of lake channels is interestingly related to the several drift sheets. Within the area occupied by the ice at the Hessle stage the channels are generally in a very perfect condition, all their contours are sharp and intact; outside this area, but still within the glaciated region, are traces of half-obliterated channels such as those in the valleys of the Dan and Deame."[1]

More recently, Bowen has used this criterion to differentiate channels apparently referable to different glacial periods in south-west Wales. He is of the opinion that within the limits of the proposed Welsh Re-advance the form of the channels indicates a later glaciation than those outside the area.

> "Channels developed on Llanvirn shales are fresh in appearance in contrast to the degraded glacial drainage channels developed on more resistant Pennant Measures in the south;"

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1. Kendall, P.F., 1916, p. 79.

whereas one would expect the reverse to be the case, the freshest channels correlating with the resistant lithologies. Therefore, Bowen concluded that,

> "The fresher channels north of the maximum extension of the Welsh Re-Advance were developed during the deglaciation of this ice. Those to the south must be referred to the deglaciation of the Main Welsh Advance."[1]

The latter is believed to be equivalent to the Older Würm of Europe, whereas the Welsh Re-advance is correlated with the Weichselian stage.[2]

Straw, working on the Lincolnshire Wolds, attempted to differentiate 'fresh' channels from 'degraded' ones on a quantitative basis. He measured the cross-profiles of some 40 channels and plotted the characteristic steepest angle derived from one profile for each. As a result, he was able to state that

> "...channels showing increased furrowing of the flanks, undulating floors, and regraded 'intake' sections rarely have sides greater than 15 degrees, and mostly between 9 and 13 degrees. Conversely, the smooth-sided, uniformly-sloping channels typically have slopes within or near to 20 to 25 degrees."[3]

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1. Bowen, D.Q., 1965, p. 344-345.
2. Op. cit. p. 330-331.
3. Straw, A., 1961, p. 118.

Straw concluded that

> "These measurements and observations are taken to be indicative of degree of freshness of a channel...",

and that the first group are older than the second.

> "It seems reasonable to assume that the overflow channels of the Wolds were created in relation to two phases of ice advance and retreat, evidence for which, (is also) drawn from the stratigraphy, disposition and morphology of the various drifts..."[1]

The two phases of ice advance were considered to have occurred during the 'Last' glaciation, separated by a true 'interstadial' period.[2]

The reference of particular meltwater channels to separate glacial phases on the basis of their state of preservation is, however, a little unsatisfactory. The 'fresh' appearance of channels is a subjective criterion, for what one person may call 'fresh', another may term 'degraded'. As yet there have been few studies in which detailed measurements on meltwater channels have been carried out; factors such as the slope of channel sides in relation to varying lithologies and the angle of the break or change of slope between the floor and the sides have rarely been measured. Until many more such studies have been made, it is perhaps premature to state that only 'fresh' channels can be referred to the last glacial episode and

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1. Op. cit. P. 119.
2. Op.cit. p. 127.

that all 'degraded' ones must be of an earlier age.

In Pembrokeshire, John has attributed meltwater channels to different glaciations on the basis of deposits found within them. He considers that the channels of Dewisland were eroded during an earlier glaciation then the Last ('Dewisland'). Glacial till, referable to the latter, overlying head has been found in some channel floors; while at Druidston and Cwm Mawr raised beach gravels lie on the rock floors of the channels at their seaward ends, indicating that some channels, at least, were cut prior to the "Poppit Interglacial" period of high sea-level.[1] John considers that this is supported by evidence of the orientation of the channels which do not coincide with the inferred ice-gradient of the Last Glaciation.[2] He concludes that,

> "...there appears to be abundant evidence that the meltwater channels of Dewisland were cut, at the latest, during the Early Glaciation."[3]

John's criterion for dating meltwater channels appears rather more satisfactory than that of Kendall and Bowen, but it also suffers a number of disadvantages. It necessitates dateable material being found within channels, whereas many which are eroded in bedrock contain post-glacial peats and hillwash. It also assumes that the channels were not re-occupied by ice which deposited till during the same glacial episode.

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1. John, B.S., 1965 (a), p. 428-429.
2. Op. cit. p. 411.
3. Op. cit. p. 429.

The fact remains, therefore, that as yet there is no satisfactory means of differentiating meltwater channels which are referable to separate glacial periods.

## SIGNIFICANCE OF MELTWATER CHANNELS

The form and position of meltwater channels allows certain conclusions to be drawn regarding the extent of the ice, the slope of the ice surface, and the character of deglaciation.

> "It is generally accepted that marginal drainage channels will only operate below the snow line, for above that level the cross-section of the land-ice contact will be concave upwards due to the marginal snow accumulation and to a more rapid outflow of the thicker ice some distance from the ice margin."[1]

Below the firn line, however, there will be a more rapid melting of ice adjacent to the valley sides, and the land-ice contact will be convex upwards (fig.4.3). In some cases a stagnant lateral zone can be seen separating the living glacier and the valley slope.[2]

Mannerfelt believed that marginal meltwater channels and benches are particularly important in deglaciation studies as they

> "...give some, event if uncertain, indication of the lowest level of the firn line and a chance of

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1. Hollingworth, S.E., 1952, p. 107.
2. Schytt, V., 1956.

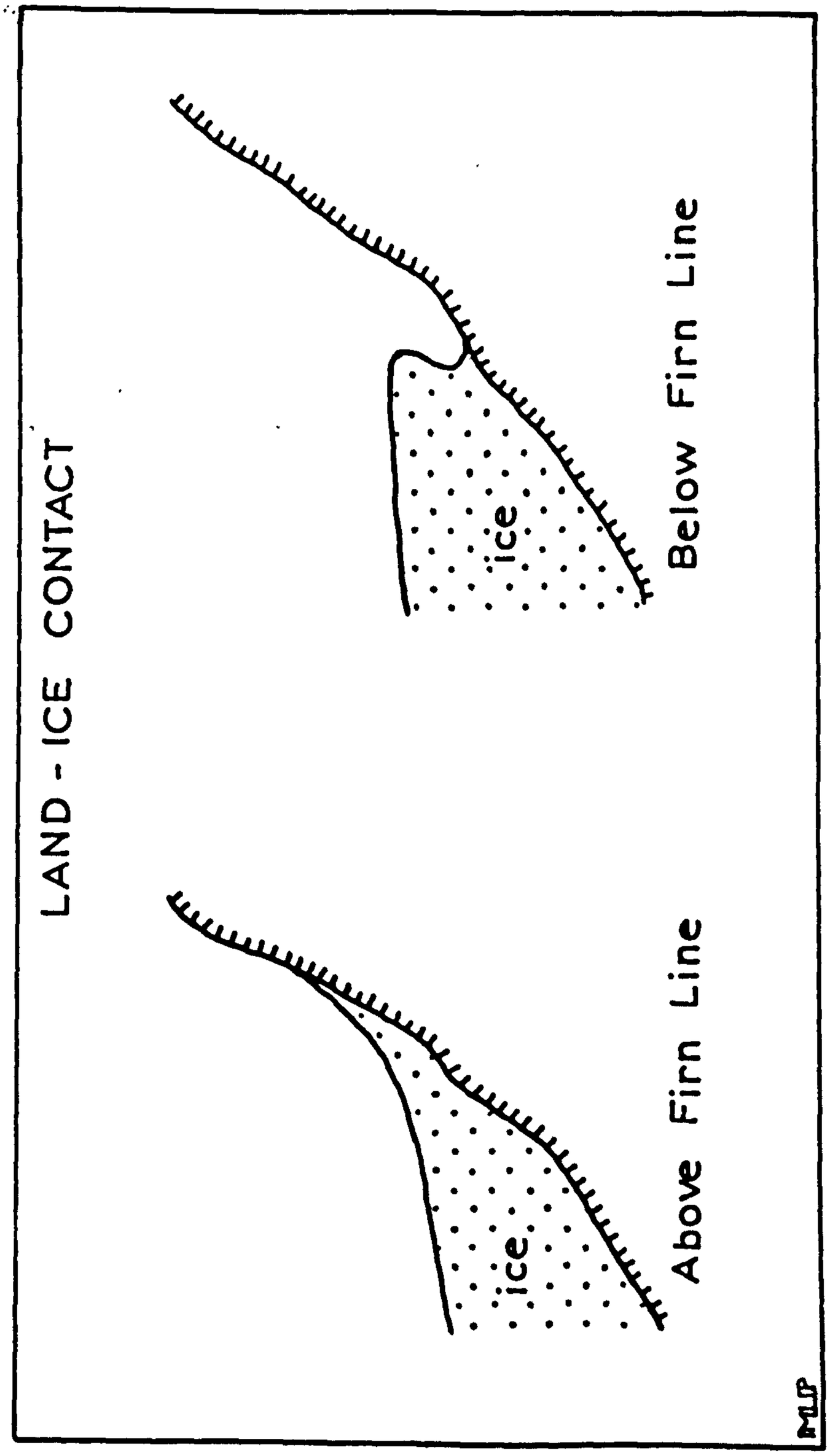


Fig. 4.3 Different types of land-ice contact

> estimating the gradient and annual melting rythm of the inland ice."[1]

He held the view that marginal channels are formed annually, mainly in the spring and early summer by the intensified melting of the snow, and that where obvious gullies occur in large numbers, they are invariably so regularly spaced that they must be assumed to reflect the annual ablation. On this basis, Mannerfelt deduced that annual thinning by ablation must have averaged 3.5 - 5 m. in certain parts of Scandinavia.[2]

Following Mannerfelt, Sissons used marginal channels, and other supposed ice-marginal features, to map the slope of the ice surface and various retreat stages during the deglaciation of part of East Lothian[3], and Embleton used similar methods to determine the slope of the ice surface in the Irish Sea basin, where "A slope of 1:200 over North West Wales would probably diminish to 1:300 or 1:400 near the Isle of Man"[4]

However, as Price stated,

> "Both the slope of the ice surface and the rate of downwasting can only be ascertained when there is no doubt that features used to determine them were essentially formed at the ice margin. With recent developments in the concept of submarginal and sub-glacial drainage, such certainty

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1. Mannerfelt, C.M., 1949, p. 194.
2. Ibid.
3. Sissons, J.B., 1958 (a).
4. Embleton, C., 1964 (b), p. 410-411.

cannot be obtained."[1]

There is also considerable doubt that marginal channels are formed annually and reflect the melting rythm of the ice. Detailed investigations have shown that the regularity in vertical spacing of marginal channels makes great variation in detail. Schytt[2] has shown that the interval between successive channels along the margins of the Moltke glacier would, on this basis, imply that the ice must have sunk some 15 - 20 m. a year, and that some streams would have had to cut down some 30 m. in the short spring flood. This is considered improbable and it is concluded that the channels cannot have been formed annually. Schytt proposed that marginal channels form in association with the development of lateral shear planes and ice ridges in the stagnant marginal zone of a glacier. As the glacier continues its retreat, a new lateral ridge will develop and a new 'inner' channel will result. The channels ought to be farther apart on a gentle slope than on a steep slope. Mannerfelt himself later accepted the possibility of a mode of formation other than annual rythm of ice melt.[3]

However, although a marginal origin for any meltwater channel is often difficult to prove, and it is only with considerable caution that these may be used to determine the actual slope of the ice surface and the rate of downwasting, the position of submarginal and subglacial drainage channels will give some indication of the

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1. Price, R.J., 1963 (b), p. 326.
2. Schytt, V., 1956.
3. Mannerfelt, C.M., 1960.

slope of the ice surface. Gjessing has argued that,

> "The direction of...subglacial drainage had two components, the one coinciding with the direction of the slope of the ice surface, the other pointing down the slope of the terrain,"

and that

> "...the subglacial pressure gradient, depended on the slope of the ice surface, determined the directions of the subglacial drainage...."[1]

Therefore, Gjessing concluded,

> "All channels, those formed at the ice margin and those formed under the ice, indicate the direction and slope of the ice margin."[2]

Assuming that all meltwater channels were formed or utilized during the last glacial episode, their distribution will give some indication of at least the minimum limits reached by the ice.

> "The criterion of marginal drainage and overflow channels as marking the limit of extraneous ice rising against ice-free ground is legitimate...."[3]

But it is quite another matter to consider that in general the highest channels mark the maximum encroachment of ice since meltwater channels are formed only below the firn-line.

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1. Gjessing, J., 1960, p. 443.
2. Op. cit. p. 446.
3. Hollingworth, S.E., 1952, p. 107.

## FORMS SIMILAR TO MELTWATER CHANNELS

Although a number of criteria have been cited which distinguish meltwater channels from abandoned valleys of normal erosion, these may sometimes be confused with similar features of a different origin.

Periglacial activity may produce forms which are akin to glacial meltwater channels. A periglacial origin has been proposed by Klatka for certain flat-floored dry valleys in the foreland of the Kysogony Mountains, Poland.[1] These resemble meltwater channels in that

> "The dry flat-floor always separated from the slopes by a distinct angle is the most characteristic feature. It normally has no runnihg water and often even no channel"

and

> "The longitudinal profile shows usually several inflexions due to layers of especially hard rock, or alluvial fans."[2]

Klatka believes the majority of the valleys to be

> "...fossil forms that originated in the late Pleistocene in the periglacial zone. At present they only revive, but are not being formed."[3]

Unlike meltwater channels, these appear to be mainly confined to

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1. Klatka, T., 1955, p. 217-219.
2. Op. cit. p. 217.
3. Op. cit. p. 218.

> "...pediments which surround residual hills and areas covered with loess, where they show together with corrasional troughs and denudational spurs a very characteristic radial pattern."[1]

Klatka does not make clear the processes which formed these valleys and is forced to admit that

> "On the basis of present data it is not possible to decide whether dry flat-floored valleys were formed as a result of corrasion, or if they are older forms modified only under periglacial conditions."[2]

Such valleys could, therefore, possibly have originated as meltwater channels during a glacial episode and be merely modified and revived by corrasion under later periglacial conditions.

Periglacial activity in the form of the transport of avalanche material may lead to chute forms which could be confused with sub-glacial chutes. The chutes are

> "...usually narrow and shallow (metres to tens of metres), in detail complicated, elongated (tens to hundreds of metres) steep gullies, or rather scars with a consequent down-slope course, often found in regular series on the upper parts of

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1. Op. cit. p. 217.
2. Op. cit. p. 218.

> (high) mountain valleys and on certain cirque walls."[1]

Matthes described such forms in the Sierra Nevada Mountains of California which he considers to be formed almost exclusively by avalanches:

> "Avalanche sculpture, wherever it is well-developed, gives the cliffs a distinct fluted appearance, there being an alternation of smoothly concave gullies and sharp rock ribs, all trending downward roughly parallel to one another in the direction of the greatest declivity."[2]

In agreement with Matthes, Rapp[3] and King[4] have expressed the opinion that

> "...avalanche erosion probably is an important morphological process."[5]

Rapp states that the process by which avalanches may erode chute forms is by the

> "Removal of frost-split debris from the ground in the track. Scratching and corroding with rock waste as tools, thus excavating avalanche chutes (gullies) even in resistant bedrock."[6]

Markgren, however, considers they may be polygenetic phenomena, partly due to fluvial erosion, partly to avalanching;

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1. Markgren, M., 1964, p. 14.
2. Matthes, M., 1938, p. 634.
3. Rapp, A., 1959, p. 46; 1961, p. 135.
4. King, C.A.M., 1959, p. 367.
5. Rapp, A., 1959, p. 34.
6. Rapp, A., 1960, p. 127.

> "The forms are polygenetic in their present state. It is considered that the initial forming was non-glacial, probably pre-interglacial. It is also suggested that a certain degree of fluvial forming preceded the development of avalanching and rockfalls."[1]

She suggests they were initially formed as a

> "...result of differential weathering, supported by slope wash, and later on, avalanching during nunatak stages."[2]

Rapp considers that avalanche chutes are found on

> "...ice slopes in and under cols...."[4]

but Markgren considers that,

> "Although it is probable that flat summits are an advantage to the evolution of chute slopes, it is not a necessafy condition. Sharp mouhtain ridges emerging just above the surface of recent wasting glaciers may be excellently furrowed by chutes."[5]

Twidale has described forms which resemble subglacial chutes in a ridge near Schefferville, Central Labrador - Ungava, and concluded that they were products of periglacial action.

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1. Markgren, M., 1964, p. 91-92.
2. Op. cit. p. 100.
3. Rapp, A., 1960, p. 157.
4. Markgren, M., 1964, p. 121.
5. Twidale, C.R., 1956 (b).

The valleys were termed "Vallons de Gelivation", and described as having a vertical backwall, and a modified V-shaped cross-profile; their long profile is irregular due to the outcropping of resistant rock strata. They possess an 'alcove-like' morphology, and at their outlets are accumulations of protalus moraine. Twidale suggested that these forms -

> "...were the product of repeated freeze-thaw cycles, leading to the disintegration of the bedrock along major joint lines aligned at right angles to the strike of the argillites. Expansion of the valleys into the ridge was probably caused by the freezing of meltwater trickling down the rock-snow interface in spring. The resultant frost-riven material was removed by solifluction. ...It was postulated that the valleys were post-glacial in age and were periglacial phenomena."[1]

Andrews, however, expressed a contrary view and considered the valleys to be products of fluvio-glacial erosion.

> "On the crest of the Dally Ridge and immediately adjacent to the 'vallons de gelivation' are beautifully preserved glacial striations, which suggest that,

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1. Quoted by Andrews, J.T., 1961. p. 3.

> if the 'vallons' are periglacial in origin, then there is a significant anomaly in the effectiveness of the processes within a few feet."[1]

Andrews considers that there is

> "...legitimate doubt that course scree material can be removed from a vallon by solifluction, for by its very definition solifluction is incapable of moving material of this grain size, unless it overlies, or incorporates particles of the sand-clay fraction."[2]

He is of the opinion that there has been very little periglacial weathering and modification accomplished since post-glacial time.

Glacial meltwater channels also occur on the ridge, often in intimate association with the 'vallons de gelivation', and Andrews concludes that

> "...the 'vallons de gelivation' are but a variety of sub-glacial channel"[3]

and proposes the name 'structually controlled chutes'.

> "It is suggested that they ('vallons') are at least earlier than the last glacial in age, and owe their initial location to the widening of joint lines by frost action. They were later modified by meltwater action...."[4]

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1. Op. cit. p. 4.
2. Ibid.
3. Op. cit. p. 7.
4. Op. cit. p. 8.

A particular difficulty in the recognition of subglacial chutes is that they may often be confused with post-glacial gullies. Sissons was aware of this and quoted a number of criteria by which the two may be distinguished. Some chutes possess a distinctive channel-like morphology, with broad flat floors and steep sides, unlike gullies which are generally V-shaped and very winding.

> "Some ...begin on the crest-lines of spurs and most, unlike normal stream valleys, are of approximately constant size throughout their lengths....In a large number of cases a channel falls steeply down the slope of a spur, but then terminates suddenly. It is inconceivable that such channels were cut by normal streams whose waters suddenly ceased to erode part way down steep slopes."[1]

Similarly Price mentions that the chutes often

> "...have abrupt beginnings and endings and do not always go straight down a slope."[2]

Often the chutes follow courses that can be explained only by postulating the presence of glacier ice when they were being cut.[3] Some head parallel to the valley side for a short distance before turning abruptly to head directly down the side to the valley floor;

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1. Sissons, J.B., 1958 (b), p. 165.
2. Price, R.J., 1960 (a), p. 88.
3. Sissons, J.B., 1963, p. 105.

unless glacier ice was present, the upper portion could not have been formed, for the stream would have plunged straight down the valley side. The distribution of the features often suggests they are subglacial chutes, rather than post-glacial gullies. Some are concentrated in the least likely place for gullying since they occur where the catchment area is very small. If the valleys form part of a complex meltwater channel system, it is reasonable to assume they are subglacial chutes, and this is often demonstrated by their relation to marginal and submarginal channels.[1]

Thus, although it is sometimes difficult to distinguish between subglacial chutes and post-glacial gullies, the morphology, position, course and distribution of the features are useful criteria.

A further problem arises in the differentiation of ice-marginal benches from similar features which are structurally controlled and formed by differential subaerial weathering and erosion. Many benches believed to be products of fluvio-glacial erosion may in fact be of structural origin. Ives and Kirby, in a critical examination of work by Derbyshire on fluvio-glacial erosion in Central Labrador, Ungava, claim that many marginal benches identified by him are, in fact,

> "...a series of structural features resulting from differential subaerial erosion."[2]

In view of this, considerable caution is required in the recognition

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1. Sissons, J.B., 1960.
2. Ives, J.D. & R.P., Kirby, 1964, p. 918.

of ice-marginal benches; consideration must be given to the detailed bedrock structures of an area, together with the possible relationships of the benches with other meltwater channels, before such an origin is postulated.

Features resembling ice-marginal benches and meltwater channels may also be produced by glacial erosion. Thus, Tarr and Martin, referring to the effects of glacial erosion in the Yakutat Bay Region, Alaska, state, that,

> "This erosion often assumes the form of a distinct grooving by which the valley walls are sculptured nearly horizontally into terrace-like steps descending gently in the direction of ice motion."[1]

These 'steps' may be confused with ice-marginal benches.

Glacial erosion may also produce features such as groove-like channels, curved winding channels, and sickle-shaped troughs (or 'sicklewannen'), termed 'plastically sculptured detail forms',[2] which often resemble meltwater channels. These features are believed to be formed by 'plastic scouring' at the base of an actively advancing glacier,[3] though there is some controversy over the actual process involved. Gjessing is of the opinion that soaked ground moraine or a viscous mixture of rock and water is the most probable agent of erosion,[4] though subglacial meltwater erosion may also be an important

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1. Tarr, R.S. & L. Martin, 1906, p. 158.
2. Dahl, R., 1965, p. 83.
3. Dahl, R., 1965; Gjessing, J., 1965.
4. Gjessing, J., Op. cit. p. 27-29.

factor. These features, however, are generally small, only a few metres deep and wide, and are usually associated with other forms of glacial erosion, such as roches moutonnée and striated rock surfaces.

# Chapter 5.

## Distribution of Meltwater Channels.

### KERRY HILLS

Black Gate col, 1368 ft., on the watershed at the head of the Mule valley, has been modified by glacial meltwaters (fig. 5.2). A system of channels, falling eastwards, is cut into the thick deposit of till at the head of the col (pl.6). Each is about 12 ft. deep, steep-sided and flat-floored, and they grade to the drift-filled valley floor below.

North of Lower Ceulannau (SO 122865) a short arcuate marginal channel is incised into the eastern side of the Mule valley, isolating a small rock knoll. Opposite the channel outlet is a small deposit of fluvio-glacial material.

Several subglacial chutes head down the valley side at Upper Wig (SO 118876). From intakes at about 1025 ft. they run for a short distance parallel to the contours, and were probably initiated by marginal or submarginal meltwaters which later turned to plunge beneath the ice.

The watershed between the Mule and Miheli valleys is cut across by a channel falling eastwards. From its intake west of Cae-betin (SO 130873) the channel pursues a sinuous course before turning abruptly towards the Miheli valley, forming a deep gorge on the valley

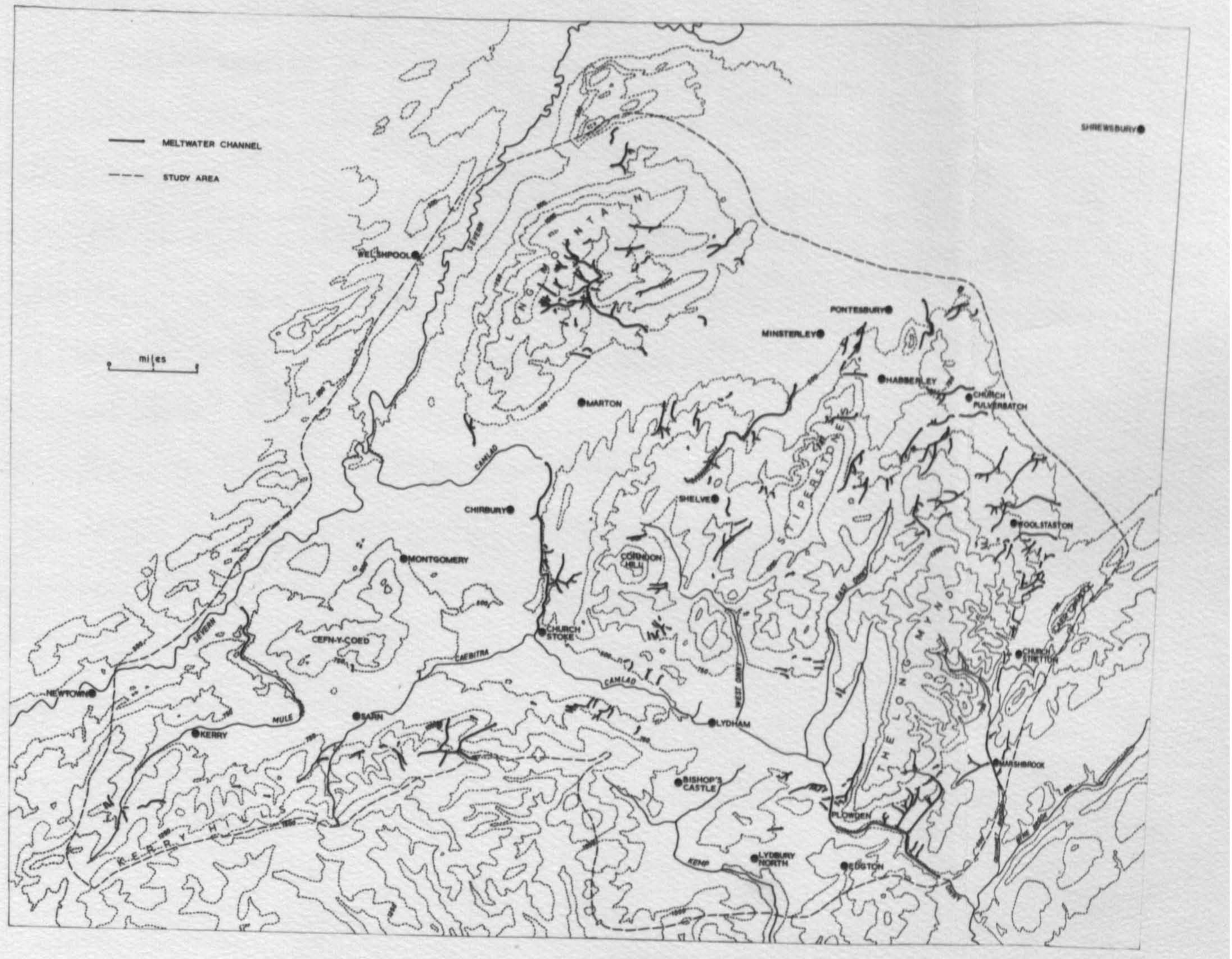


Fig. 5.1 Distribution of meltwater channels

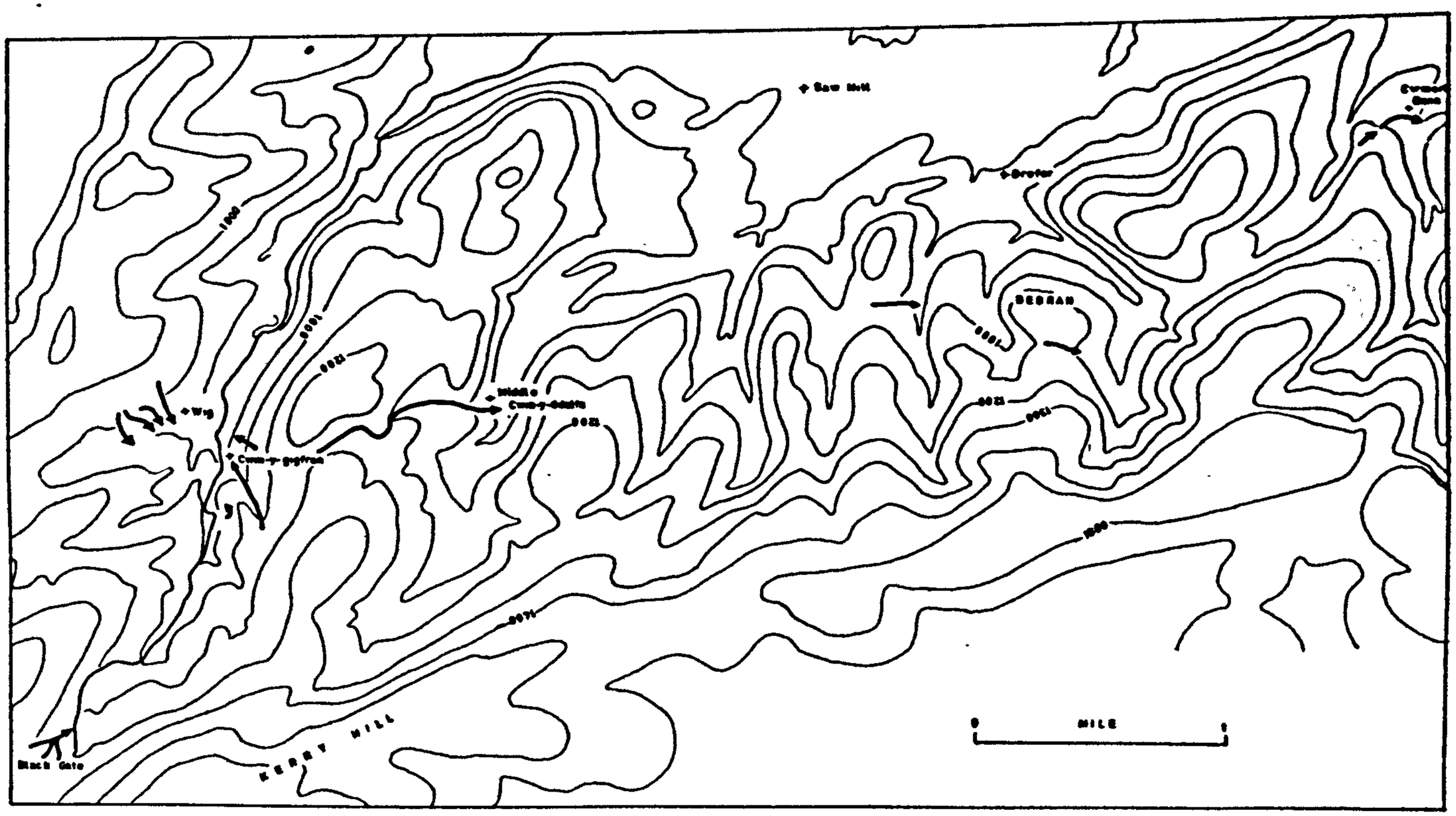


Fig. 5.2 Meltwater channels on Kerry Hills

side. It cuts across the drift terrace at Middle Cwm-y-ddalfa and its outlet is some 10 ft. above the valley floor.

The interfluve on the western side of the upper Caebitra valley is cut across by two meltwater channels. A marginal bench, the intake of which is at 950 ft., heads eastwards diagonally across the interfluve at a gradient of about $1^{o}$. After a few yards it is continued by a channel which turns abruptly to plunge down the side of the Caebitra valley as a subglacial chute.

A well-developed system of meltwater channels is situated on the hillslopes above and to the east of City (SO 211899) (fig. 5.3). Two channels, 1[1], with intakes at 1400 ft., heads northwards across the plateau surface of the Kerry Hills towards the head of Bahaillon Bank dingle. It is probable that they were eroded subglacially by meltwaters plunging into the valley head. At the lower end of the dingle a marginal channel is incised some 25 ft. into the valley side; it is open at both ends, the intake and outlet hanging above the valley floor. North of the valley, channel 2 is cut into the side of Pant Hill at 1200 ft.

A shallow col at the southern end of Pant Hill is occupied by channel 3, the intake of which is at 1325 ft. and hangs above the side of Bahaillon Bank dingle. It was probably eroded by meltwaters flowing eastwards, away from the dingle. At its outlet the channel makes a sharp turn towards the north, into the head of Cwm Hopton. That the meltwaters continued eastward at a high elevation is indicated

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1. Numbers refer to channels on fig. 5.3.

by channel 4. The intake is above the eastern side of Cwm Hopton, directly in line with Channel 3. The channel heads eastwards towards Pant Glas at an average height of about 1300 ft. and with a slope of about 2° for 1¼ miles. It cuts across the watershed of the Kerry Hills and runs into the head of a tributary to the Upper Unk Valley. (Pl.13). A number of smaller channels head northwards across the plateau into channel 4. The largest of these is channel 5 the intake of which is at 1425 ft. Two small channels, 6, head parallel to the southern side of channel 4 before turning abruptly to plunge down its side as chutes.

The low gradient of channel 4 and the fact that it heads sub-parallel to the contours suggests it was eroded close to the ice margin. The tributary channels on its southern side, including the two marginal channels, 6, indicate the presence of ice above channel 4 during at least part of its excavation. Thus the channel was probably cut submarginally. The incision of the channel across the watershed at Pant Glas was probably accomplished by the superimposition of subglacial meltwaters. Subsequently, it appears that meltwaters plunged directly beneath the ice, down the steep slopes of the Kerry Hills, cutting two prominent subglacial chutes. The incision of the chute 7, and the continued flow of meltwaters along part of channel 4 has led to an overdeepening of the floor of the latter, which is now about 25 ft. below that section heading towards Pant Glas. The chute, 8, is incised up to 75 ft. into the floor of a pre-glacial valley. Again, the continued flow of meltwater along

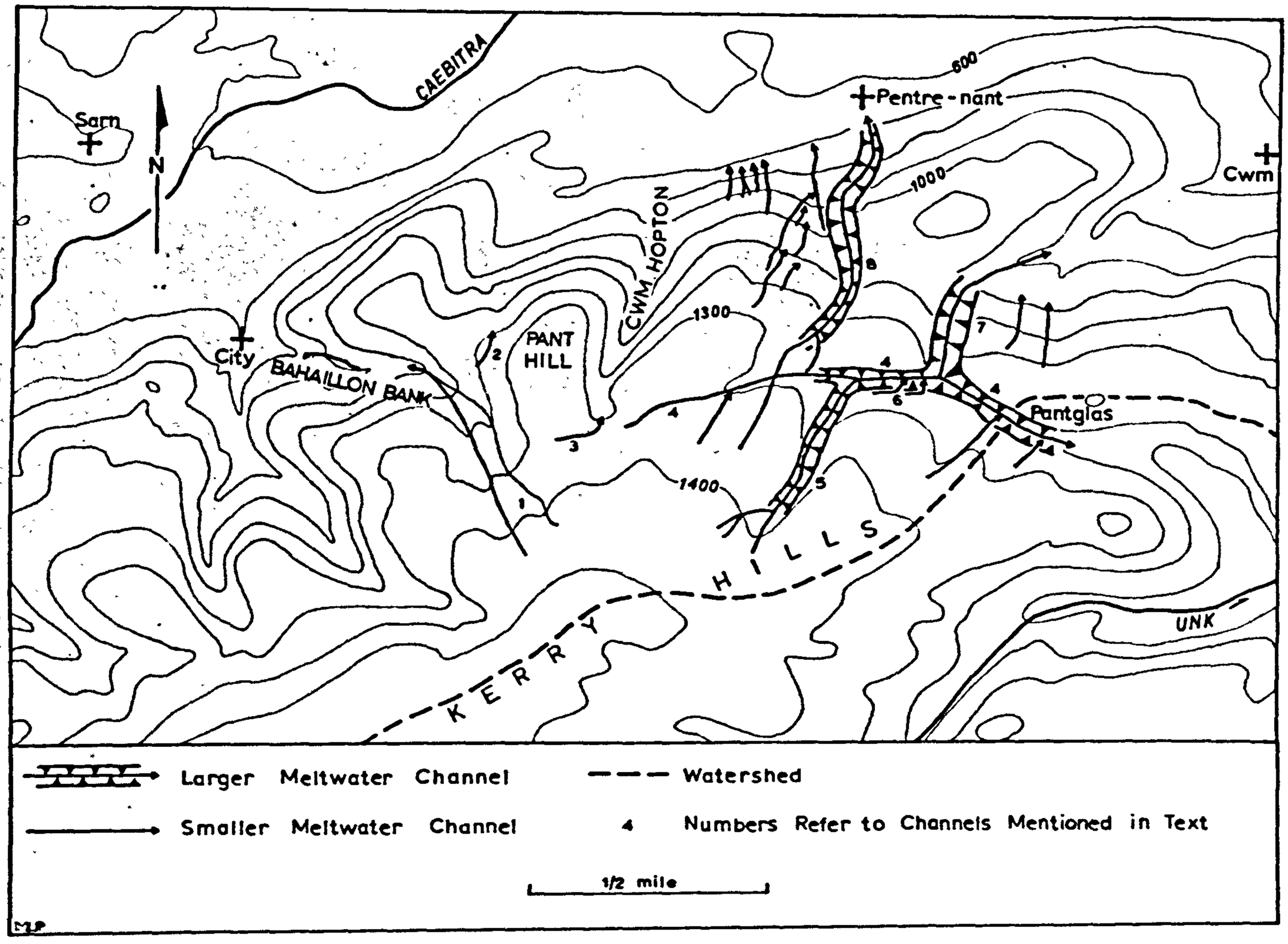


Fig. 5.3 Meltwater channels at City, Kerry Hills

Pl. 12 Kame deposits in the Vale of Stretton

Pl. 13 Meltwater channel above City, Kerry Hills

the upper part of channel 4 has led to its overdeepening.

Immediately west of channel 8, a group of discontinuous, meandering chutes heads directly down the hillside. They are generally less than 25 ft. deep and have a gradient of about 14°. The intake of the highest is at 1275 ft. and they consistently begin and end part-way down the hillslopes.

## CEFN - Y - COED HILLS

The lower portion of the river Mule flows through a prominent gorge, the origin of which is important in an explanation of the diversion of the river out of the Mule-Caebitra valley and towards the Severn. The question of drainage diversion will be discussed in a later chapter.

The construction of generalised contours (fig. 5.4), gives some idea of the form of the landscape before dissection. It is evident that there are two large hill masses reaching to over 1000 ft. on either side of the Mule, and that between them is a low-lying area, generally below 700 ft. Also, it appears that there was a pre-glacial watershed across the present Mule at Waen.

The gorge section of the Mule valley begins at Upper Maenllwyd from whence it deepens rapidly and pursues a winding course towards the Severn at Abermule. The gorge is narrow and steep-sided, and in its upper section it is markedly asymmetrical, the eastern side being considerably higher and steeper than the western. This is particularly

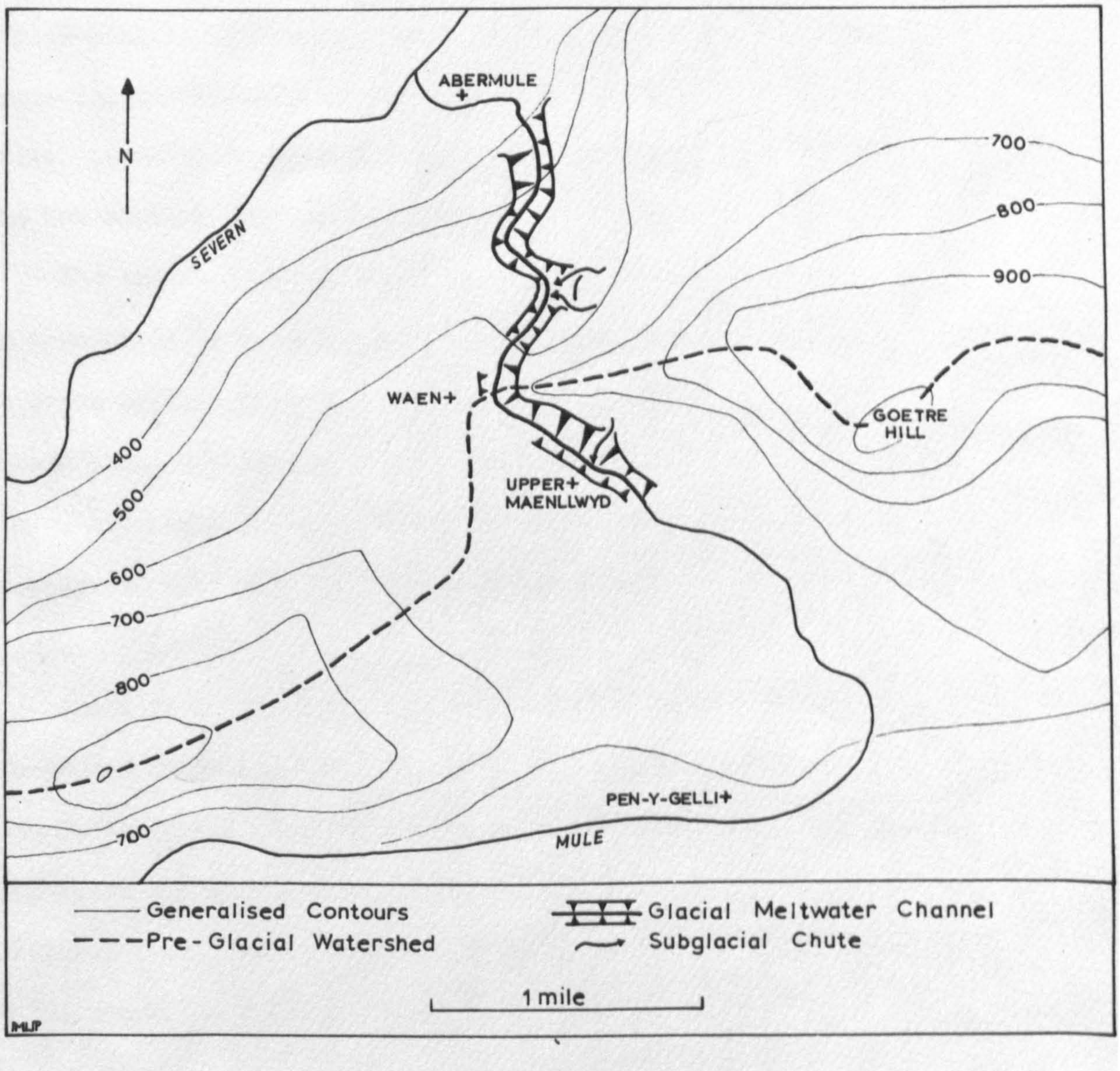


Fig. 5.4 Relationship between the Mule Gorge and topography

evident at Waen [illegible], where it is incised across the pre-glacial watershed. Not only does the gorge cut directly across the relief, more or less at right angles, but also truncates several ice-moulded ridges. Several chute-like valleys plunge directly down the eastern side of the gorge.

The incision of the gorge across the pre-glacial watershed, its truncation of several spurs and ridges, the marked asymmetry of its cross profile, and the presence of chute-like valleys on its eastern side, combine to suggest the gorge was cut by glacial meltwaters. It seems likely that it was eroded subglacially, being superimposed from the ice across the watershed and the underlying relief.

Apart from the Mule gorge, the only other channels on the Cefn-y-Coed hills include a few subglacial chutes. For example, above Cwmbellan (SO 202924) a channel heads parallel to the contours for a short distance before turning abruptly to plunge down the hillside. The channel intake is at 850 ft., and its outlet is about 100 ft. above the valley floor.

## MONTGOMERY - CHIRBURY GAP

### Marrington Dingle

On the eastern side of the broad depression underlain by Wenlock shales, separating the Cefn-y-Coed and Corndon Hill masses, is one of the largest and most interesting meltwater channels in the Welsh

Borderlands, renowned among Palæozoic geologists as Marrington Dingle (fig. 5.5).

This narrow, sharply cut gorge heads northwards along the edge of the higher Ordovician country, from its intake at 425 ft. at Church Stoke for some 4 miles to its outlet at 300 ft. at Hockleton Bridge. The cutting of the gorge has resulted in a remarkable diversion of drainage: the combined waters of the Camlad and Caebitra rivers, meeting at Church Stoke, are channelled northwards across a prominent watershed and into the Severn drainage system. The incision of Marrington Dingle across a watershed has had a peculiar effect:

> "In following the thickly wooded banks of this little stream from Church Stoke to the north-west, the traveller, who merely hears the gurgling of the water, is impressed with the belief that he is ascending to its source, while in reality he is descending towards its mouth. This deception is produced by the partial rise of the hills towards the northwest, though the bottom of the valley really deepens in that direction."[1]

At its intake at Church Stoke the gorge is incised some 15-20 ft. into a broad, drift-covered platform overlooking the Camlad valley, which can be traced for some distance into the dingle. (Pl.14). Two small dry channels plunge down the eastern side of the dingle

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1. Murchison, R.I., 1839, footnote p. 551.

from this platform.

Northwards, the dingle deepens rapidly as it cuts into the rising hill-mass. Between Calcot and Church Stoke the dingle has a composite cross-profile, indicating that it has been incised into the floor of a pre-existing valley which drained towards the south (fig. 8.10). At Calcot, a number of valleys are tributary to the eastern side of the gorge, the largest of which is Coed Brook. These drain southward and westward, and in their lower parts they also have a composite cross-profile, indicating a deep gorge incised into the floor of a pre-existing valley. This factor, and the angle at which they enter the dingle, suggests that they originally formed the headwaters of the southward-draining valley between Calcot and Church Stoke.

Immediately to the north of Calcot the dingle makes a sharp turn to the west, breaching a former watershed formed by a narrow ridge of Hagley Volcanics. At this point the cross-profile of the gorge is asymmetrical, the eastern side standing 50 ft. above the western (Pl.15 ). The asymmetry of Marrington Dingle becomes more pronounced towards the north. The ground surface into which the gorge has been cut slopes away to the west, and the eastern side stands high above the slope.

Opposite Marrington, a number of meltwater channels plunge down the eastern side of the dingle. One of these (1)[1], heads northwards

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1. Numbers refer to channels indicated on fig. 5.5.

Pl. 14 Entrance to Marrington Dingle, Church Stoke

Pl. 15 Marrington Dingle, immediately below the prominent bend

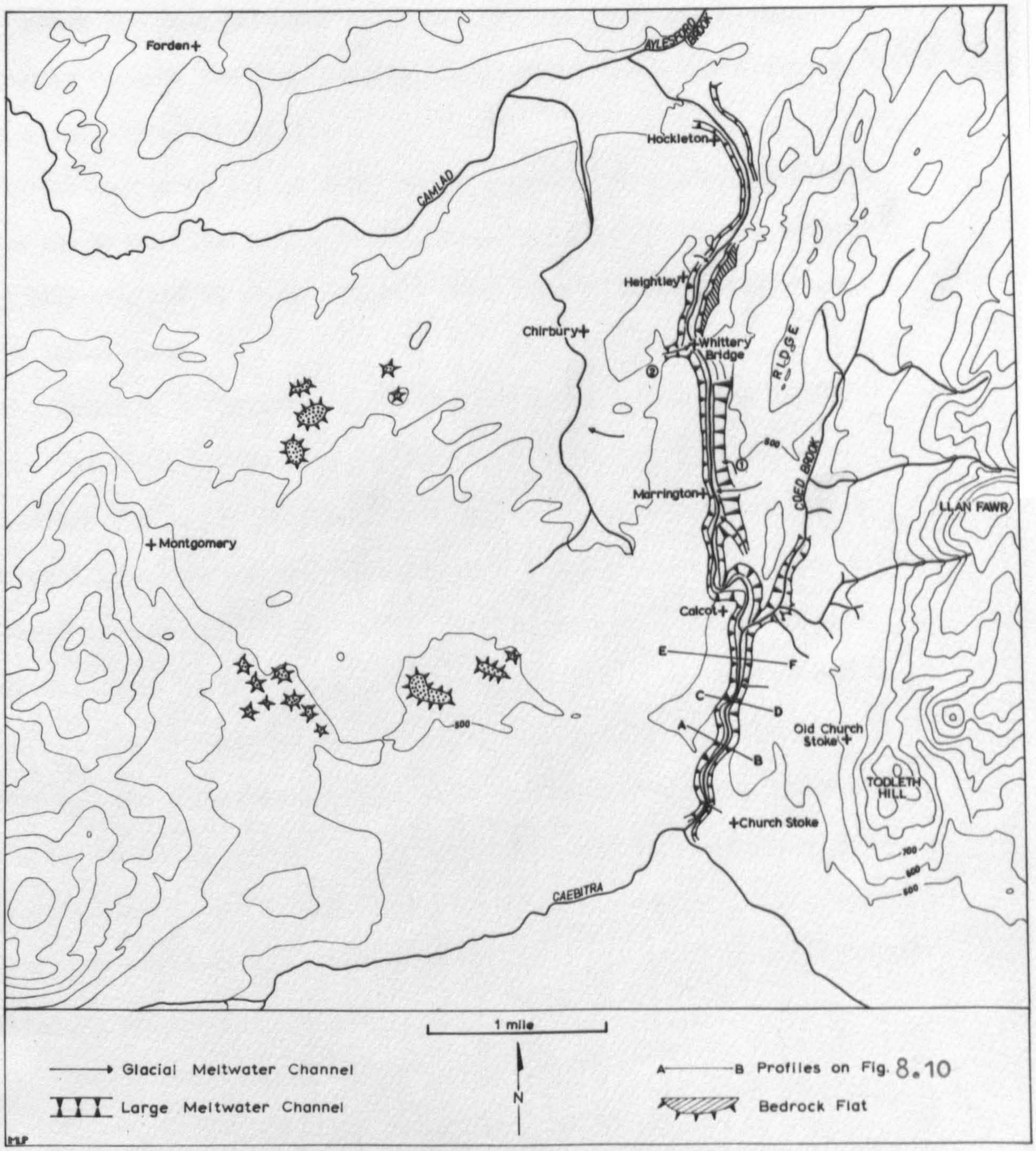


Fig. 5.5 Marrington Dingle

for about 100 yd., parallel to the valley side, before turning abruptly towards the dingle; its outlet hangs above the floor of the dingle by about 100 ft.

At Whitterey Bridge a meltwater channel (2) enters the dingle from the west. The intake is at the point of confluence of several dry valleys, and it seems probable that these directed meltwaters into the channel.

Opposite Heightley a broad bedrock bench stands some 15 ft. above the floor of Marrington Dingle, and is backed by a steep hillslope ($25^{o}$). It is probable that this represents two stages in the erosion of the gorge: an early phase when a one-sided channel was incised into the hillside, and a later phase when a gorge was cut into the floor of the former one-sided channel.

At its outlet, the dingle is incised some 20 ft. into a broad drift-covered platform standing above the Marton Valley. Although this is underlain by bedrock, it is morphologically continuous with the extensive spread of glacial drift extending across the Marton valley, on the surface of which are kettle-holes and a number of kames.

<u>Origin of Marrington Dingle</u>

Harmer suggested that Marrington Dingle,

> " ...seems to have originated as an overflow channel to a small glacial lake,"[1]

situated in the depression south of Church Stoke. It was envisaged that

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1. Harmer, F.W., 1907, p. 481.

the area north of Church Stoke was blocked by an ice stream and that

> " ...the drainage of the region south of that place was forced against the higher Ordovician land, probably flowing at first between it and the ice, but eventually excavating a channel through the former ...."[1]

This explanation of Marrington Dingle was later accepted by Dwerryhouse and Miller.[2]

Several factors, however, suggest that this is not the correct explanation. There is no evidence that the area south of Church Stoke was the site of a glacial lake. The brickearth cited by Harmer as a lake deposit is, in fact, glacial till[3]. In addition, there are numerous kettle-holes in the area indicating that the site of the supposed lake was occupied by stagnant ice.

A peculiar feature of the channel is that for the first mile it occupies the floor of a valley which formerly drained in the opposite direction before turning sharply to cross a prominent watershed. This is inexplicable in terms of overflow from a glacial lake. A more likely explanation is that glacial meltwaters flowing northwards, on or in the ice, were let down across the watershed and incised into it. They were perhaps deflected by the prominent ridge, and maintained this course with continued downcutting, thus explaining the bend in

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1. Ibid.
2. Dwerryhouse, A.R., and A.A. Miller, 1930, p. 102.
3. See p. 102

the channel. With continued incision, the meltwaters took advantage of the pre-existing valley and were superimposed into its floor. The overdeepening of Coed Brook and its tributaries is probably also due to erosion by subglacial meltwaters utilising the pre-existing valleys.

North of the watershed, the marked asymmetry of the cross-profile suggests the channel was formed by marginal or submarginal meltwaters cutting into the rising ground to the west. Initially, the meltwaters would probably excavate a one-sided channel, as at Heightley, but subsequent erosion would produce a two-sided feature, the western side standing high above that to the east.

Both north and south of the watershed several chutes plunge down the western side of the channel. These provide further evidence for a subglacial origin of Marrington Dingle as they can only be accounted for if glacier ice existed on its upslope side when it was being formed. Furthermore, the course of channel 1, and the fact that its outlet hangs above the dingle floor, suggests it was eroded by marginal or submarginal meltwaters which subsequently turned to plunge into ice occupying the present site of Marrington Dingle. The meltwater channel which enters the eastern side of the dingle at Whittery Bridge was also probably eroded subglacially.

It therefore seems likely that Marrington Dingle was formed submarginal to an ice-mass occupying the low ground between Montgomery and the high Ordovician country to the east. Englacial meltwaters were superimposed across the former watershed at Calcot and the channel was

eroded sufficiently deep to cause a permanent diversion of drainage towards the north.

Since the direction of the subglacial drainage was in direct opposition to the slope of the ground surface to the south of the watershed, the direction of drainage must have been determined by the slope of the ice surface. This is analogous to a situation described by Gjessing in Sweden:

> " ...the subglacial pressure gradient, dependent on the slope of the ice surface, determined the directions of the subglacial drainage, all subglacial water in the area between the culmination zone and the present watershed ... was forced to ascend to a pass in the present watershed on its way to the valleys on the other side."[1]

Thus the slope of the ice surface in this area during deglaciation was probably down towards the north.

## LONG MOUNTAIN

Glacial meltwater channels on Long Mountain (fig. 5.6) constitute two principal types with regard to position: those which are closely related to the topography, which include channels located within broad valley heads and deep gorges eroded in the floors of pre-existing valleys; and secondly, channels which are anomolous in position with regard to topography, which include those eroded into

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1. Gjessing, J., 1960, p. 443.

the sides of larger channels or valleys, and those which sever spurs, cols and interfluves. The former type predominate, and there are comparatively few examples of channels which are anomolous in position.

The channels occur right up to the main watershed of the Long Mountain, up to a maximum elevation of 1275 ft., and are concentrated on a relatively gentle south-east and east side of the mountain.

Complex systems of anastomosing channels are located in several valley heads immediately to the east of the Long Mountain watershed. At Beacon Ring a system of dry, shallow, winding channels, 1,[1] is located in a broad valley head (Pl.16 ). The intakes are at about 1200 ft., and the channels are 10-15 ft. deep. Some extend only part-way down the hillslope, but most converge to the valley floor around Mulsop (SJ 276059). A number of meltwater channels are also incised into the broad valley-head of Outrack Dingle. One channel, 5, heads southeastwards into the valley from its intake at 1275 ft. on the Long Mountain watershed. A system of dry, shallow, winding channels, 7, is located in the valley-head west of Rhos Farm (SJ 283073) (Pl.17). At the latter, a gorge some 15 ft. deep has been incised into the pre-existing valley floor, and the valley side has been breached by the intake of channel 8 which heads directly down the hillslope into the Rowley Brook valley.

Two prominent channels 9, are located in the broad valley-head southeast of Welsh Harp (SJ 279080). The intake of that to the south is at 1150 ft., and the channel is winding, V-shaped and some

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1. Numbers refer to channels indicated on fig. 5.6.

Pl. 16 Meltwater channels located in a broad valley head close to the Long Mountain watershed

Pl. 17 Meltwater channel at Rhos Farm, Long Mountain

10-15 ft. deep (Pl.18 ). At its outlet, the channel grades to the floor of that to the north. The intake of the latter is also at 1150 ft., but the channel is up to 100 ft. deep, steep-sided, and has a broad flat floor. Small chutes plunge down the sides of the upper part of the channel, and the outlet of one hangs some 50 ft. above the channel floor. At Upper Monksfields (SJ 285076) a smaller channel has been cut into its northern side. The intake hangs above the channel floor by 25 ft., but its outlet grades to the floor; it is markedly asymmetrical in cross-profile.

Beyond the point of convergence of the valley-head meltwater channels, the pre-existing valleys typically show a valley-in-valley cross-profile. Meltwaters following the course of the valleys have incised gorges, up to 25 ft. deep, into their floors. These all head into the Rowley Brook valley.

Anomolous meltwater channels are also situated in these valleys. Northeast of Mulsop a prominent channel, 2, severs the interfluve between the Mulsop valley and Outrack Dingle. (Pl. 19 ). The intake hangs above the floor of the Mulsop valley by about 100 ft., but its outlet grades to the floor of Outrack Dingle. Lower down the Mulsop valley a large channel, 3, has been eroded into its side. The intake hangs above the overdeepened valley floor by about 75 ft., but its outlet grades to the floor. The channel is markedly asymmetrical in cross-profile, the northern side being over 30 ft. high, compared with only 15 ft. on the southern side.

The Rowley Brook valley, which in its upper portion contains the Rowley brook and lower down the Brockton brook, is a broad, deep

Pl. 18 Meltwater channel in a valley head on the Long Mountain

Pl. 19 Meltwater channel cutting across a spur on Long Mountain

feature which heads generally southwards towards the Marton valley. The upper portion of the valley is unmodified by fluvio-glacial erosion, and a small stream flows over the broad, drift-covered valley floor. Below about 875 ft. the floor has been overdeepened and steepened, a modification which increases downvalley. At Pound House (SJ 292070) a gorge has been incised some 30 ft. into the broad valley floor. In places, one side of the valley has been considerably steepened by fluvio-glacial erosion, so that the valley resembles a one-sided channel. North of Middle Walton (SJ 292055), the eastern valley side, Tantree Bank, has been cliffed and steepened to a height of over 100 ft. Lower downvalley glacial meltwaters have eroded a gorge, 40-50 ft. deep, termed Beechfield Dingle, down the sides of which plunge small subglacial chutes. This eventually opens out into the broad Rowley valley which enters the Marton valley at Brockton (SJ 309053).

East of the Rowley Brook valley fluvio-glacial erosion has exerted a similar influence on the landscape: some meltwater channels are anomolous in position, but in general they are closely related to the topography and are located in the floors of valleys.

South of Plantation Cottage (SJ 304092) a meltwater channel, 14, has been incised into the floor of a broad valley heading eastwards. A number of subglacial chutes plunge down the southern side of the channel, the largest of which is that at the Forest of Hayes (SJ 086310).

To the south, channel, 15, Breakneck Dingle, is incised into the floor of a broad, eastward-heading valley. The intake, at 1075 ft., is in the floor of a broad col in the watershed of the Long Mountain. Modification of both the above valleys by glacial meltwaters ends at about 800 ft., below which are broad, drift-filled valleys.

At Rowley (SJ 306066), a gorge, 16, has been eroded into the floor of a valley trending eastwards along the margin of Aston Hill. The gorge is, in places, up to 50 ft. deep, steep-sided, and has a broad,flat floor over which flows a small stream. A number of smaller channels head down both sides of the gorge, and generally grade to its floor. One runs parallel to the side of the gorge for a short distance, before turning to head obliquely down it.

The lower part of the Wallop valley has similarly been modified by glacial meltwaters. A large gorge, Causebeech, 17, is incised into the valley floor, and at Whitton Grange (SJ 337084) a number of chutes plunge down its northwestern side.

In addition to the meltwater channels occurring within the principal river basins of the Long Mountain, a number of channels are situated on the margins of the hill.

East of Forden (SJ 228008), a gorge, 18, has been eroded into the floor of the lower part of a broad valley heading southwards to join the Marton valley at Rhyd-y-groes (SJ 248007). Two smaller channels are tributary to its eastern side.

A number of gullies, 19, head down the steep edge of the Long Mountain below Marton Crest. Many of these have the appearance of

channels or chutes. Some head obliquely down the hillslope and then turn abruptly to plunge at a high angle. They characteristically begin and end part-way down the hillslope.

To the north, two channels, 20, termed Flying dingle and Yell dingle are situated on the lower slopes of the Long Mountain. The headwaters of these are directly aligned with each other, separated by a shallow depression, below which both turn abruptly to plunge down the hillslope as chutes. It seems probable that meltwaters originally flowed from Flying dingle to Yell dingle, but were later diverted to plunge directly beneath the ice.

West of Brockton, Ragbeech dingle, 21, has the appearance of a meltwater channel. It has a 'double-intake' at 450 ft., and is 15-20 ft. deep and steep-sided. Its outlet at 400 ft. hangs above the floor of the Marton valley.

To the northeast are two similar channels, Workhouse dingle, 22, and Hawcocks dingle, 23, each 20 ft. deep and steep-sided. They begin suddenly at the confluence of shallow depressions between ice-moulded ridges, and head directly down the hillside, cutting across the ridges.

Origin of the Channels

The complex systems of anastomosing channels located in broad valley heads below the watershed of the Long Mountain are probably of subglacial origin. Evidence for such an origin is the fact that they often run obliquely, instead of directly down the hillside, and where they meet they often show discordant relations to one another; they are sometimes discontinuous, or begin and end part-way down the

hillslope. Furthermore, the chutes on the sides of channel 9, and the fact that the outlet of one of these hangs above the channel floor, suggest a subglacial origin. The asymmetrical channel incised into the side of channel 9 must have been eroded marginally or submarginally to ice occupying the latter.

Meltwater channels occupying valley heads are usually explained by the superimposition of englacial meltwaters onto the underlying topography. A broad embayment type of valley head, as found on the Long Mountain, allows latitude for the superimposition of englacial streams as complex channel systems; several streams becoming deeply incised independently before uniting as the valley narrows downstream.

Below the valley heads, meltwaters were concentrated in the valley floors, so that channels are intimately associated with drainage lines. Their very position suggests they were cut beneath the ice, which is also suggested by the chutes on the channel sides.

In some cases, meltwaters flowed marginally to ice occupying the valleys, eroding one-sided channels, as at Tantree Bank in the Rowley Brook valley.

The few channels which are anomolous in position on the Long Mountain are also probably subglacial. Channel 2, which cuts across the interfluve between the Mulsop valley and Outrack dingle, was probably formed by the superimposition of an englacial stream onto the underlying spur. The asymmetry of the cross-profile of the channel suggests that it was eroded submarginally to an ice-mass,

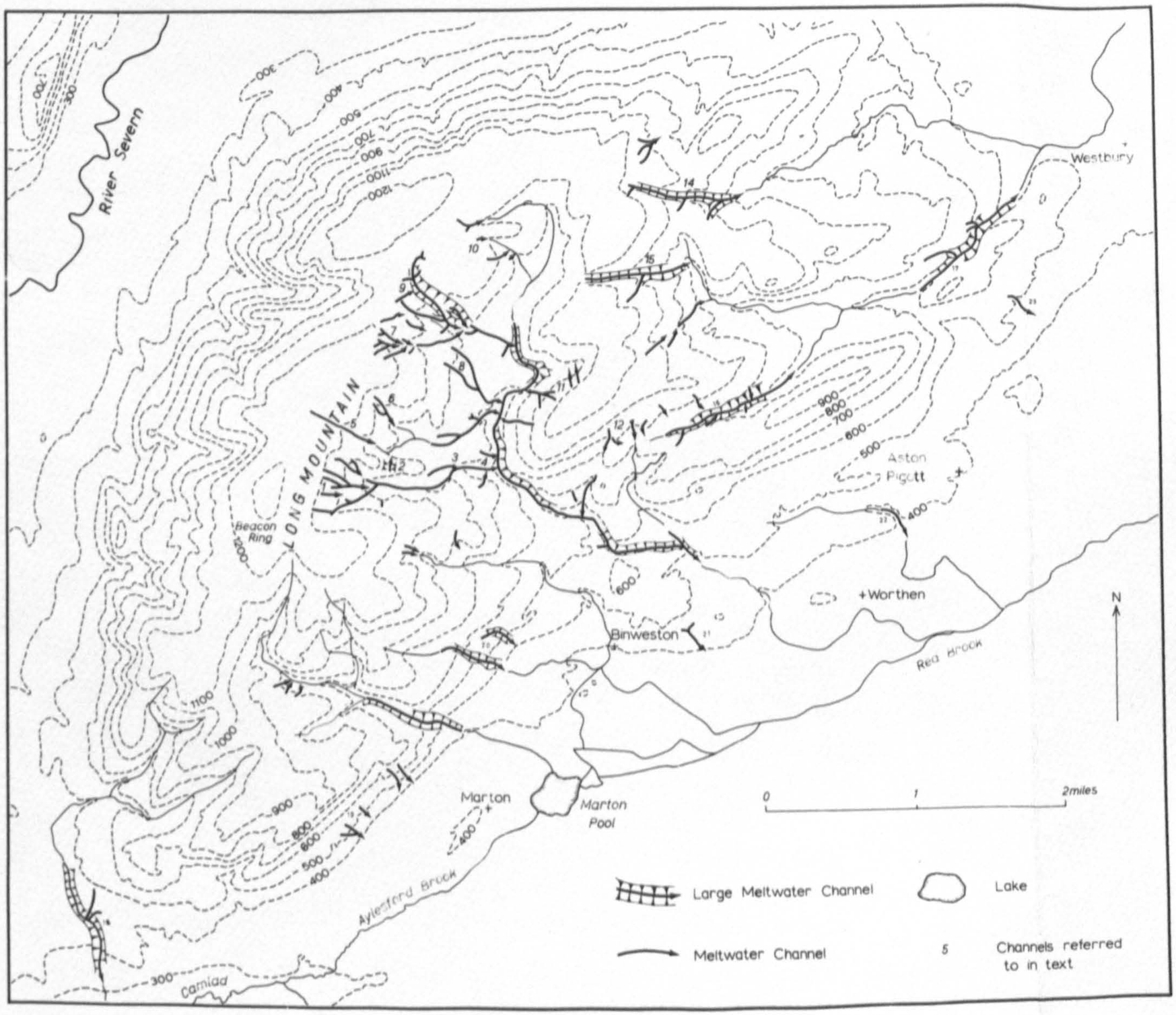


Fig. 5.6 Distribution of channels on Long Mountain

and the fact that its intake hangs well above the floor of the Mulsop valley indicates that ice occupied the latter at the time of erosion.

Similarly, the incision of channel 15 across the watershed of the Long mountain suggests it was eroded by the superimposition of englacial meltwaters into the floor of the pre-existing col.

## MIDDLETOWN GAP

The broad gap separating the Long Mountain from the Breidden Hills is occupied by two large meltwater channels, heading from east to west. A group of hills, rising up to 550 ft., form a prominent watershed across the gap at Middletown; both channels are incised across this watershed.

At Breidden Station one channel heads first southwards before turning abruptly towards the west, cutting across the watershed at Glyn Common. The channel is up to 50 ft. deep, narrow and steep-sided, and numerous chutes plunge down its eastern side; the outlets of the latter hang about 10 ft. above the channel floor.

Another, smaller, channel is situated to the west of Middletown. This begins as a one-sided feature, heading westwards; it then turns abruptly south and again towards the west, cutting across the watershed. It seems probable that both of these channels are of subglacial origin, formed by meltwaters being superimposed from the ice across the watershed. The chute plunging down the side of the larger channel lends support to this proposition.

## CAMLAD - ONNY VALLEY

On the lower slopes of Aston Hill, at Upper Aston (SO 289914), a series of benches have been cut into the hillside, (fig. 5.7). They occur as a group, one below the other, between 775 ft. and 700 ft., and slope down towards the west at about 3°. To the west, at Lower Bank (SO 269916), another narrow bench at 775 ft. can be traced along the contours for about 100 yd. Closely associated with these benches are a series of steep-sided gullies which run down the hillside at a high angle. The course followed by some of the gullies strongly suggests they were cut by glacial meltwaters flowing subglacially. The largest is Aston Dingle, which runs for a short distance parallel to the contours before turning abruptly to descend the hillside. East of Middle Aston (SO 294913), the gully turns to head sub-parallel to the contours for about 200 yd., and then reverts to its straight descent. To the east and in direct line with the oblique section of the channel, a shallow, dry gully extends for 50 yd. before turning to fall into the main valley below.

The intake of Aston Dingle was probably cut by marginal or submarginal meltwaters, which soon found their way directly into the ice, eroding a subglacial chute. It seems likely that the oblique section of the channel, at Middle Aston, was cut submarginally, at a later phase of deglaciation, the meltwaters later plunging directly into the ice isolating the oblique channel to the east. In view of their close association with the gullies, it is probable that the

benches were also eroded by glacial meltwaters. Their gradients suggest either a marginal or submarginal origin.

On the northern side of the Camlad valley, between Todleth Hill and Roveries Wood, are a number of small channels. West of Snead (SO 318921), between 550-600 ft.,there are several which run oblique to the contours. Associated with these are gullies which run directly down the hillside, the intakes of some of which run parallel to the contours for a short distance. The gullies were probably eroded subglacially, and the oblique channels submarginally.

At Hyssington (SO 314940) there is a complex system of meltwater channels. They generally run from north-west to south-east, cutting across spurs, and then turn towards the south. Their almost complete independence of the topography strongly suggests that they were eroded by the superimposition of englacial meltwaters.

East of the watershed at Lyndham numerous channels, cut into both bedrock and glacial drift, appear to have directed meltwater drainage eastwards towards the Plowden Gap at the southern end of the Long Mynd.

On the lower slopes of Norbury Hill several channels have been cut sub-parallel to the contours. A small channel, the intake of which is at 900 ft., runs for a short distance parallel to the

large valley which drains southwards past Norbury; it then turns abruptly towards the east, its outlet hanging above the valley floor by about 75 ft. On the opposite side of the valley, and in direct line with channel 1, another channel, 2, heads eastwards parallel to the contours, cutting across a spur. The channel is steep-sided and flat-floored and has a prominent 'hump' in its long profile. The outlet grades to a flat at 875 ft., but in direct line with the outlet is a further channel, 3, which also cuts across a spur and heads towards the East Onny valley.

It is likely that these three channels were eroded contemporaneously, by marginal or submarginal meltwaters flowing towards the east. Since the latter two are incised across spurs, and one has a 'humped' profile, they were probably formed by the superimposition of englacial meltwaters onto the underlying topography.

At a lower elevation on Norbury Hill are a number of other meltwater channels. Above Norbury, two shallow channels, 4, head obliquely down the hillside for a short distance, one which has a chute cut into its side, suggesting they were cut submarginally. At Norbury two channels, at 750 ft. and 775 ft., head eastwards parallel to the contours, and at their outlets hang above the Norbury valley by about 20 ft. South of channel 3, two further channels, with intakes at 825 ft. and 725 ft., also head eastwards, cutting across a spur. On the opposite side of the East Onny valley, the spur (SO 382923) south of Wentnor is incised by two east-west channels at heights of 750 ft. and 730 ft. respectively, and their outlets hang well above

the Criftin Brook. The higher channel has a double intake. Several small chutes plunge down the sides of the channels, and a small bedrock knoll is isolated in the floor of the northernmost one. This strongly suggests that the channels were formed submarginally, by the superimposition of englacial meltwaters across the spur of the hill. The channels at Norbury and Wentnor were formed by meltwaters flowing eastwards, submarginal to ice occupying the low ground to the south. The lower part of the East Onny valley was probably filled by ice at this time, meltwaters flowing across it from Walkmill towards Wentnor. It is difficult to say whether the succession of channels at lower elevations on the hillside represent various phases in the downwasting of the ice, since none can be said to have been formed strictly at the ice margin.

The extensive drift-covered area between the West Onny and the Long Mynd is dissected by systems of meltwater channels, now mainly abandoned. To the south of Norbury, two large channels run southeastwards towards the East Onny valley. They are about 20 ft. deep, eroded entirely in glacial till, and are separated by flat- topped interfluves.

At Myndtown (SO 390895), there are three large channels running southwards parallel to the Long Mynd escarpment, with occasional linking channels approximately at right angles. The most easterly channel is the only one at present occupied by a stream. At its northern end, west of Handless (SO 392904), the channel intake is at just over 700 ft., and the intakes of the other two channels are

about 20 ft. and 50 ft. lower. Small lakes and marshy hollows occur in the higher parts of the three main channels, perhaps due to the local persistence of stagnant ice after the retreat of the main glacier. The easternmost channel runs past Myndtown before turning sharply to the west and continuing as a 30 ft. deep chute into the Onny valley. Two smaller chutes plunge towards the Onny valley at the south-western end of the Long Mynd.

Since they form an interconnecting system of sub-parallel valleys, and since the easternmost is terminated by a chute, it is likely that the channel system around Myndtown, and that south of Norbury, was excavated in a sub-glacial environment. They were probably cut at a fairly late stage of deglaciation, beneath relatively thin ice, when there was a substantial amount of subglacial drainage flowing towards the southern end of the Long Mynd. The marshy hollows which occur in several channels are kettle holes, indicating that the channels were formed beneath the ice.

A system of subglacial channels runs eastwards down the hillsides on the western side of the Onny valley, at the entrance to the Plowden gorge. The main channel, the intake of which is at 700 ft. in the floor of the col above Friesland (SO 366887), heads eastwards through the Eaton Plantation. It is about 25 ft. deep, steep-sided and flat-floored. Several chutes plunge down both sides of the channel, one of which has a double intake. Other chutes run into the Plowden gorge near Choulton (SO 377882), from intakes between 550 and 675 ft. These channels, together with those on the opposite side

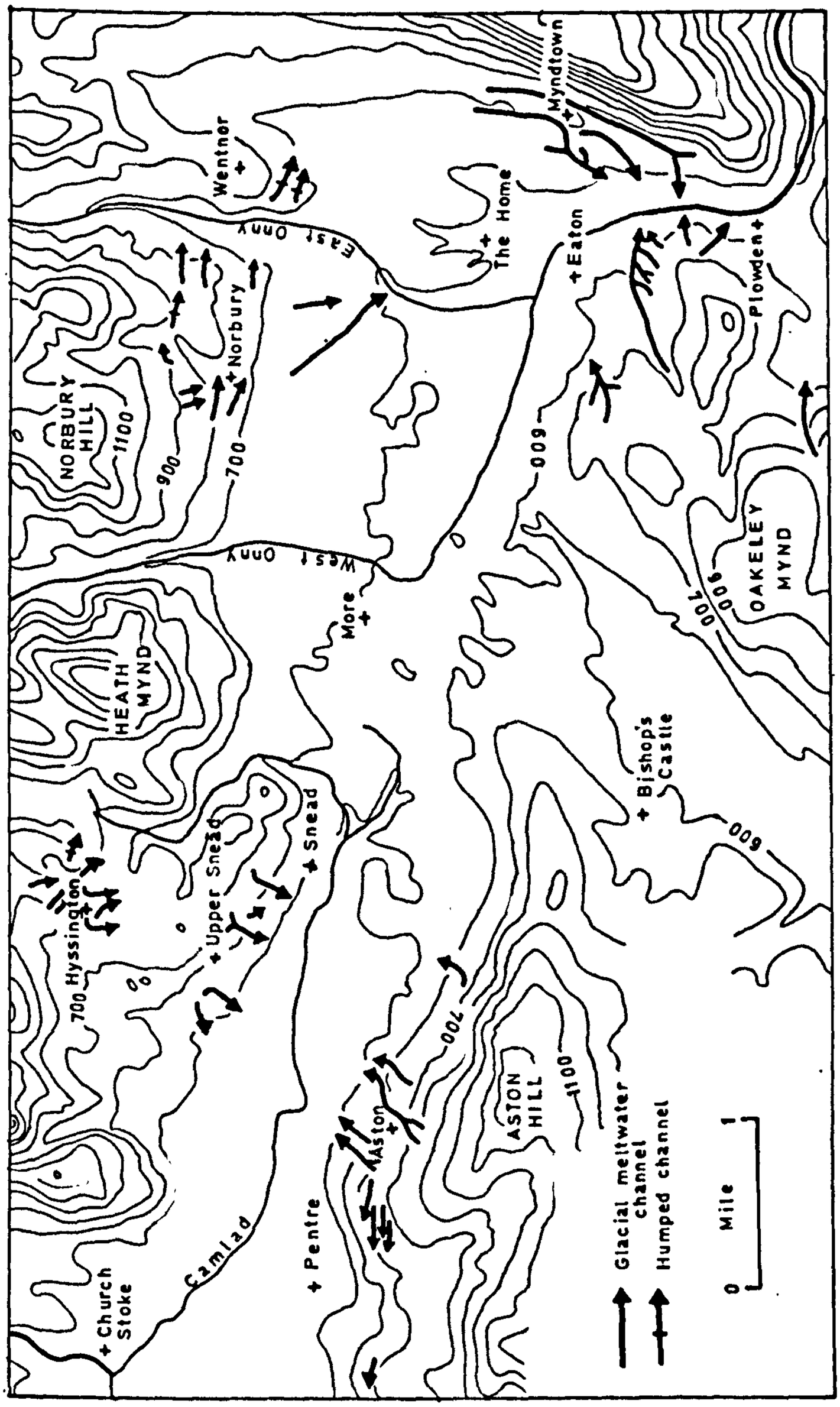


Fig. 5.7 Meltwater channels in the Camlad-Onny valley

of the Onny, run down the sides of the large channel which heads eastwards through the Plowden gap. They were formed by meltwaters plunging beneath stagnant ice at the southern end of the Long Mynd and within the Plowden gap.

## PLOWDEN GAP

Between Plowden and Horderley, the river Onny flows through a broad gap between the Long Mynd plateau and the Silurian hills to the south (fig. 5.8). On both sides of the gap is a broad bedrock platform, the surface of which is at about 600 ft. It is most continuous on the southern side, below Plowden Woods, and occurs as a discontinuous feature on the north. Elongate shallow depressions, at about 650 ft., occur in places on the southern end of the Long Mynd (SO 398877) about 700 yd. west-south-west of Hillend Farm. These features trend approximately north-east and possibly mark the course of drainage marginal to ice filling the Onny valley to the south.

The present Onny flows in a broad gorge, incised some 25 ft. below the platform. The steep sides of the gorge can be traced continuously from a point 700 yd. S.E of Eaton (SO 378890), across the entrance to the broad valley between Plowden and Lydbury North, into the Plowden gap, past Horderley towards Wistanstow. Several small channels run off the 600 ft. platform to the Onny valley. These occur, for example, at the Mill (SO 384871) and to the east of Hillend Farm. At Horderley, several meltwater channels running

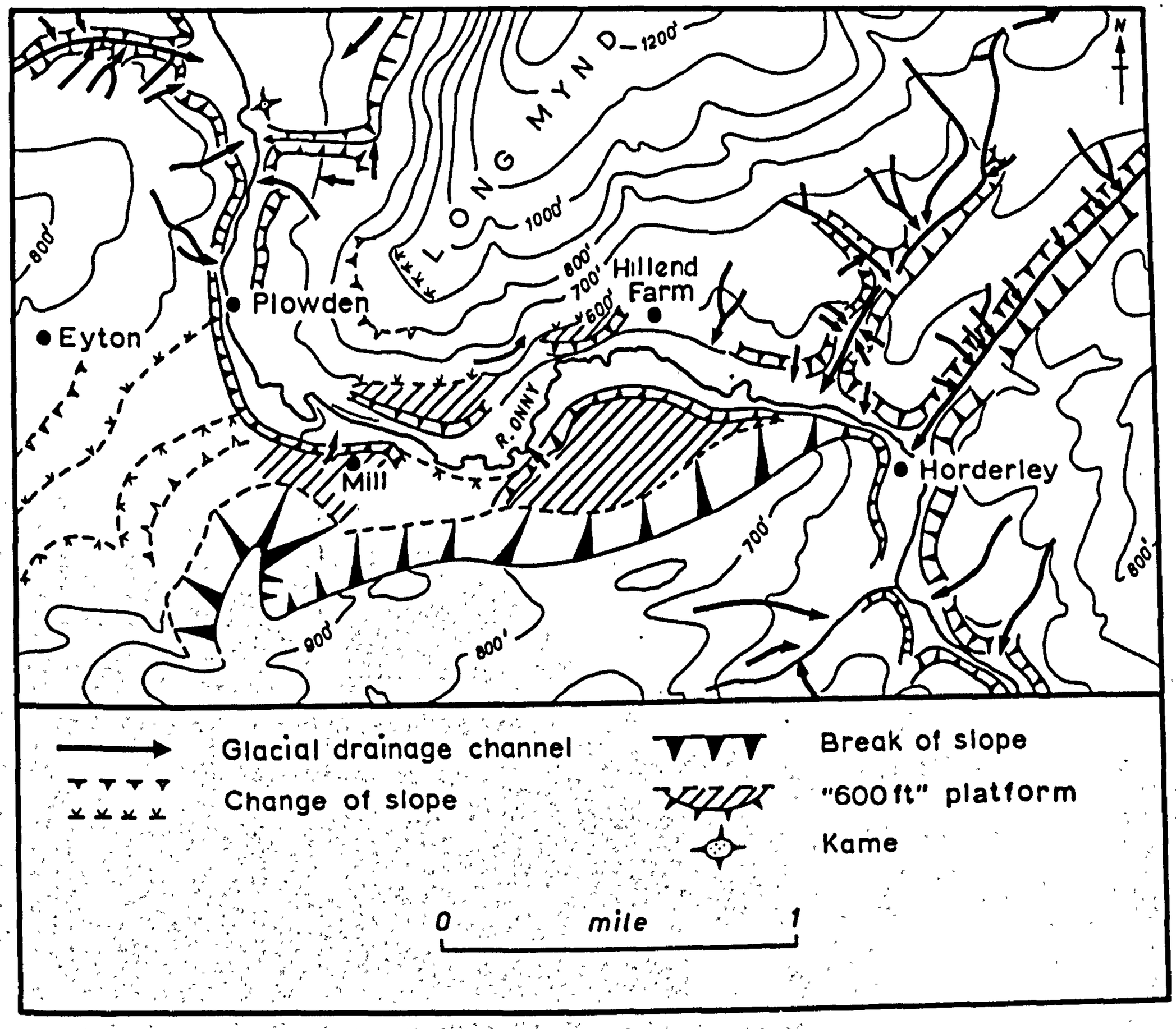


Fig. 5.8 Plowden gorge

south-westwards from the Cwm Head area enter the Plowden Gorge. Further south, two further chutes plunge into the gorge. The close association of the gorge with numerous meltwater channels suggests that it was itself excavated by melt streams. The several chutes which plunge down its sides were formed by meltwaters plunging into stagnant ice within the gorge, suggesting that it was formed subglacially by meltwaters becoming incised below the 600 ft. platform. A considerable volume of meltwaters, from both the upper Onny basin and the Cwm Head area, flowed into the gorge and were directed down the Onny valley towards Wistanstow.

## MARSHBROOK - HORDERLEY GAP

A broad valley between the Long Mynd and Whittingslow Ridge trends NNE-SSW between Marshbrook and Horderley (fig. 5.9). Two small streams at present drain the valley, flowing in opposite directions from a watershed at 750 ft. near Cwm Head.

At Hamperley two channels, forming a discontinuous valley, run northeastwards across a spur at 800 ft. One has a 'humped' long profile, suggesting they were formed by the superimposition of englacial meltwaters across the spur. A deep channel runs south-eastwards from the centre of this discontinuous valley towards Cwm Head, where it turns sharply to the southwest and runs along the edge of the Whittingslow Ridge into the Plowden gorge. Several chutes plunge down the northwest side of this channel, the intakes

of some running parallel to the channel side for a short distance. The chutes were formed by subglacial meltwaters plunging into stagnant ice within the main channel, at a fairly late stage of deglaciation, and indicate that it was excavated subglacially.

Another large channel, the intake of which is adjacent to that of the discontinuous channel at Hamperley, heads south-westwards into the Plowden gorge. It receives several tributary channels from the lower slopes of the Long Mynd, forming a remarkable incision at Churchmoor Rough. Several small chutes run down the sides of the main channel, indicating that it was also formed subglacially.

From the watershed at Cwm Head a channel runs north-eastwards into the Marshbrook gorge, and is joined at New House (SO 436899) by a channel running from a point south of Minton. Chutes plunge down both sides of the tributary channel, indicating a subglacial origin.

During deglaciation the Marshbrook-Horderley gap was apparently filled with stagnating ice, and originally it would appear meltwater drainage was directed towards the northeast, possibly reflecting the slope of the ice margin. With continued downwasting of the ice the topographic control of the meltwater drainage increased, and they flowed in opposite directions from a watershed near Cwm Head. South-westward flowing waters eroded two large sub-parallel channels which run into the Plowden gorge; northeastward flowing waters eroded the channel between Cwm Head and Marshbrook. Subglacial chutes on the sides of the channels indicate they were formed beneath the ice.

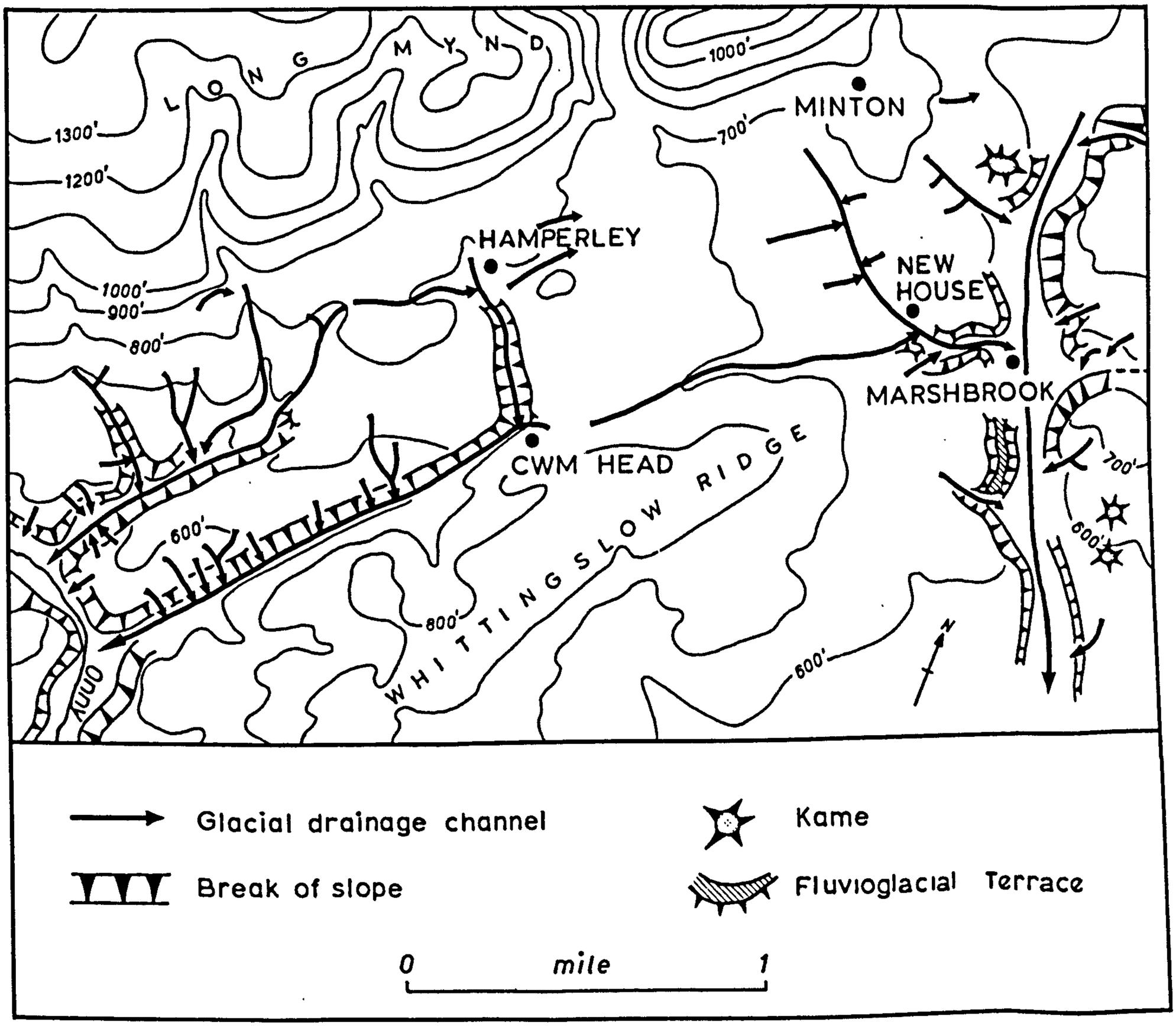


Fig. 5.9 Channels in the Marshbrook - Horderley gap

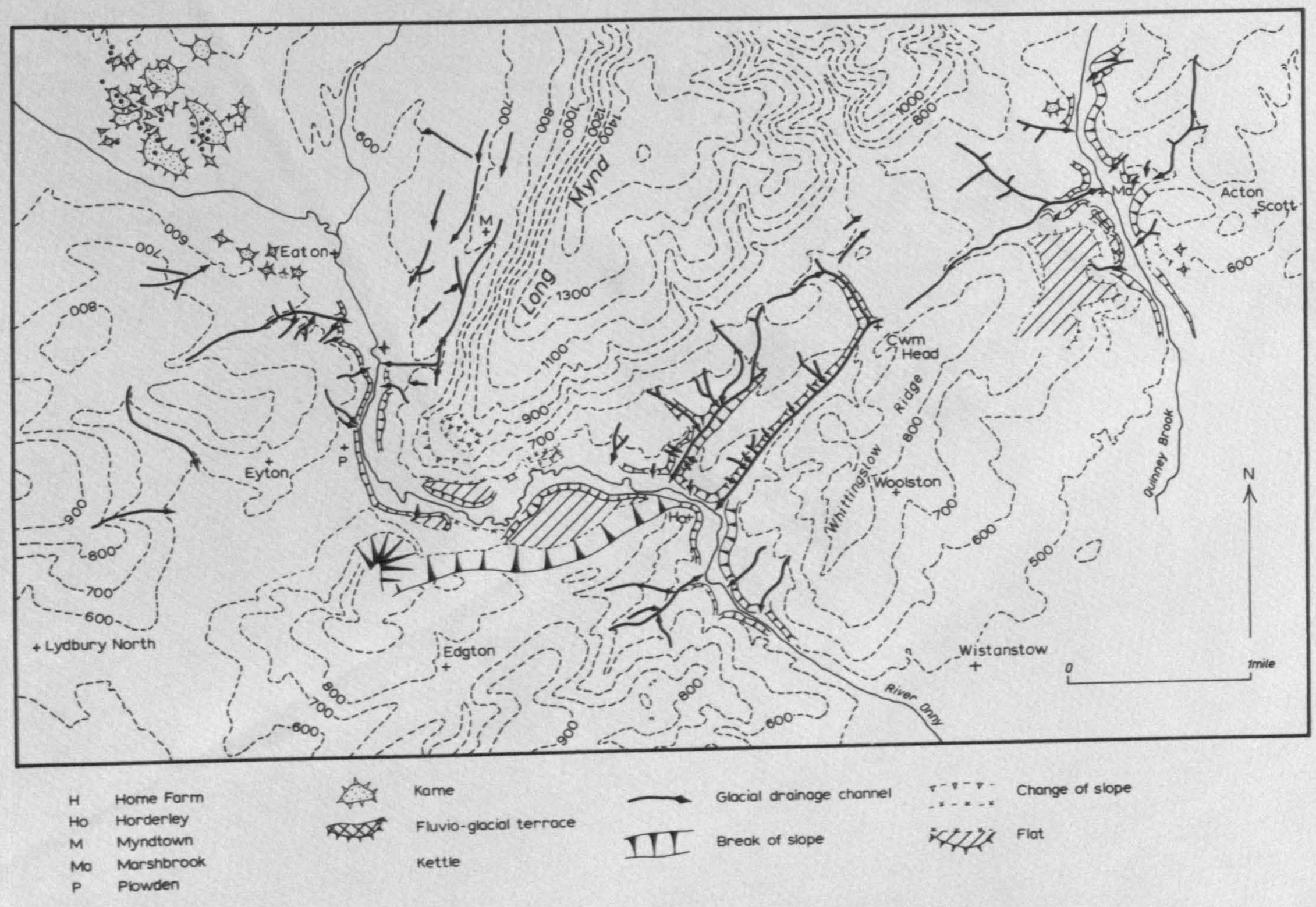


Fig. 5.10 Meltwater channels at the southern end of the Long Mynd

Wright[1] maintains that the two large channels which run towards the Plowden gorge were eroded by melt streams issuing from a glacier snout standing on the col north of Hamperley. Whittard[2] suggested that the Cwm Head channel was excavated by spill from a lake ponded up by ice standing to the north in the Church Stretton valley. Both proposals are untenable. There is no evidence for such a lake in the Vale of Stretton, which would also require a confining barrier at Marshbrook. The chutes on the channel sides indicate they were formed subglacially, not proglacially. The channels in the gap are tributary to the Marshbrook gorge and the Plowden gorge, both of which were also formed beneath stagnant ice.

## SHELVE DISTRICT

Fluvio-glacial erosion has added significant detail to the morphology of the Shelve District.

Numerous meltwater channels head in opposite directions, away from an east-west watershed to the north of Shelve village (SO 336990) (fig. 5.10). Several have intakes at about 1050 ft., one of which is in the floor of a col in the watershed.

A prominent channel head south-westwards past Ritton Castle (SO 345977), roughly along the axis of the Ritton Castle syncline (Pl.2D). The intake is at 1025 ft., to the east of Ritton Farm

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1. Wright, J.E., in Greig, D.C., and others, 1968, p. 289.
2. Whittard, J.W., 1952, p. 148; 1953, p. 244.

(SO 348982), below which the channel makes an abrupt turn to the west before assuming its south-westerly course. Above Ritton Castle the channel is 15 ft. deep, steep-sided and flat floored. Thereafter it runs along the base of Brooks Hill; its south-east side is about 10 ft. high compared with only 25 ft. on the north-west. Above the south-east side a narrow bench has been cut into the edge of Brooks Hill and several chutes plunge from this down the sides of the channel. The channel outlet is west of Lower Brookshill (SO 338968), but its course is continued by a broad valley heading towards the West Onny. At Brook House (SO 338964) a short channel, some 25 ft. deep, has been incised into the western side of this valley, isolating a flat-topped knoll in the valley floor. It seems likely that this was eroded by the same meltwaters which cut the Ritton Castle channel.

The asymmetry of the Ritton Castle channel suggests it was eroded in an ice-marginal position, while the chutes plunging down the upslope side indicate it was cut submarginally. The small channel further downvalley, at Brook House, was possibly formed by the superimposition of meltwaters into the valley side. Several meltwater channels are located on the lower flanks of Corndon Hill. On the north, a system of channels with intakes between 1050 ft. and 1150 ft. heads northeastwards, roughly parallel to the hillside. On the southern side of the hill another channel system heads eastwards into the Shelve District.

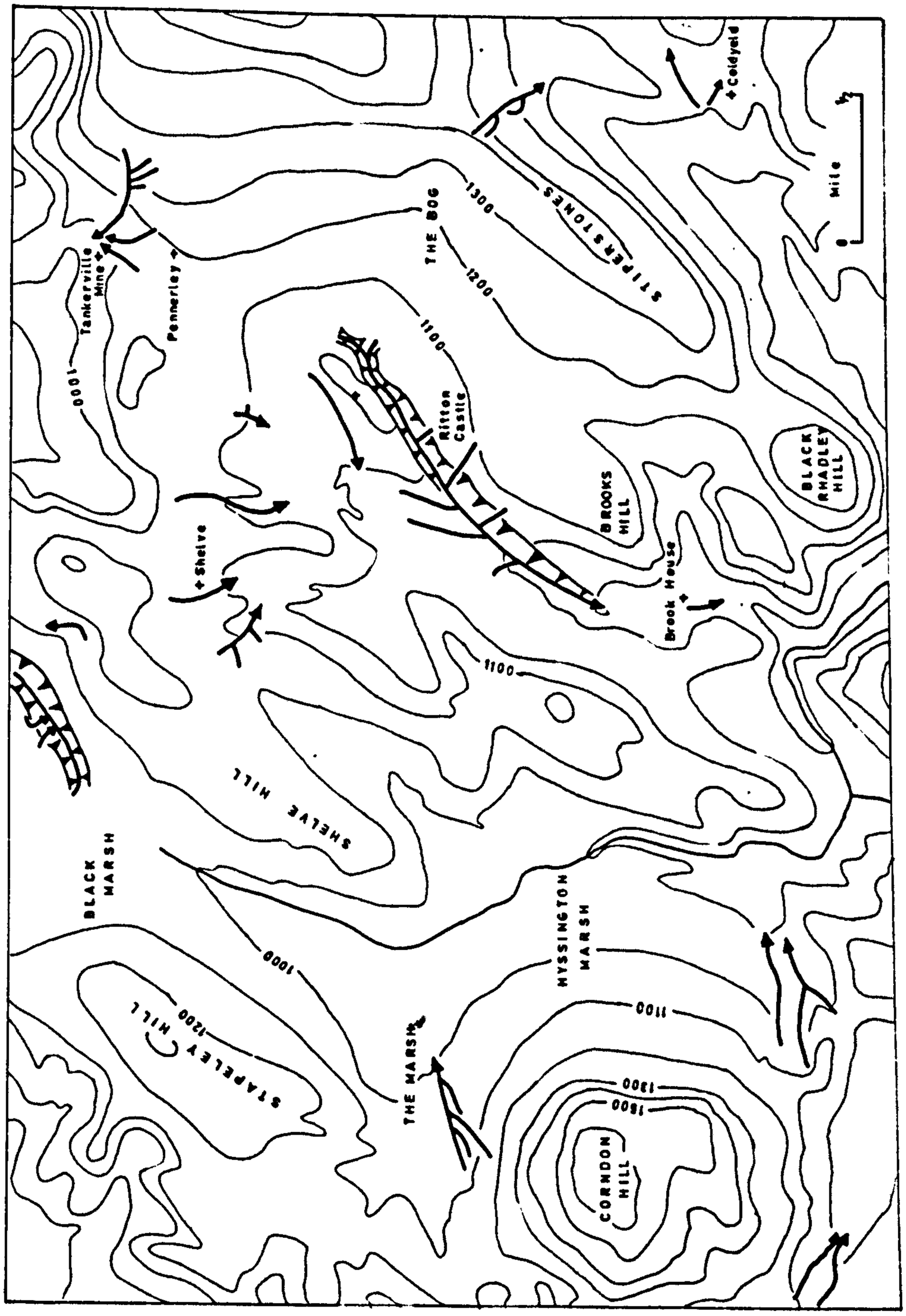


Fig. 5.11 Channels in the Shelve district

Hope valley

The largest channel which heads northwards from the watershed at Shelve is the Hope valley (fig. 5.12). This feature, which is over 4 miles long and in places over 100 ft. deep, compares in size and form with Marrington Dingle, but unlike the latter its formation did not result in any major diversion of the pre-glacial drainage.

The intake of the channel is located at Black Marsh, from whence it heads north-eastwards for about 1½ miles towards Hope along the geological contact between Ordovician shales and Mytton Flags. A prominent ridge of intrusive volcanics stands above its eastern margin. At its intake the channel has a typical U shape, with steep sides and a broad flat floor and is about 20 ft. deep. Downvalley it becomes progressively broader and deeper; in places it is 100 ft. deep, and at Hope the floor is about 100 yd. wide. Several small chutes plunge down both sides of the channel; some head for a short distance parallel to the channel side before turning abruptly to plunge to its floor.

At Hope the channel makes an abrupt turn to the south-east, cutting across the volcanic ridge formerly on its eastern margin. (Pl. 2a). At this point the channel floor is very narrow; it is markedly asymmetrical in cross profile, and appears to have been partly cut into the floor of the section above Hope. The channel then makes a second abrupt turn and resumes its original north-easterly course. The volcanic ridge through which the channel is incised at Hope now forms its northern margin for a short distance.

Pl. 20 Ritton Castle meltwater channel

Pl. 21 Hope valley, showing the bend in the channel where it crosses a volcanic ridge. Note the pronounced asymmetry of the channel sides

Again, the cross-profile is asymmetrical; the northern margin is very steep ($33^{o}$) and between 75 and 100 ft. high, while the opposite side, cut partly into glacial deposits, is much less steep ($15^{o}$) and only about 25 ft. high.

At Fox Inn (SJ 350017) the channel receives a tributary from the south. From its intake at Venusbank Farm (SJ 353013) it heads along the base of another volcanic ridge and is asymmetrical in cross profile. Several chutes, cut into drift, plunge down the southern side of the channel.

East of Snailbeach (SJ 375022) the Hope valley channel turns again and heads northwards towards Ploxgreen (SJ 367037). This section of the channel is eroded almost entirely into glacial deposits, and opens out in an area of kame deposition.

The relationship of the Hope valley channel to the pre-glacial relief is well illustrated by the construction of generalised contours (fig. 5.12). Above Hope the channel does not occupy a pre-existing valley, but is incised into a bench at about 1000 ft., marginal to the narrow volcanic ridge. A broad valley lies to the west of the channel and becomes tributary to it at Hope. It is interesting to speculate why the meltwaters did not follow this valley and erode a channel in it. Beyond the point at which it traverses the ridge, it appears that the channel follows a pre-existing depression towards Ploxgreen.

The morphology of the Hope valley meltwater channel is indicative of a complex origin. The presence of chutes on both sides of the upper section strongly suggest that this was eroded subglacially. That these are subglacial chutes, rather than post-glacial gullies, is demonstrated by the fact that several follow courses that can be explained only by postulating the presence of glacier ice when they were being cut. Those that head parallel to the channel side for a short distance were formed by marginal or sub-marginal meltwaters, which soon turned to plunge beneath the ice. Others, that head obliquely down the channel sides were probably cut submarginally. The upper part of the channel, therefore, was eroded subglacially, the meltwaters possibly picking out the geological weakness along the line of contact between the shales and flags. They would also tend to be concentrated along the break of slope below the steep volcanic ridge. These two factors would probably largely account for the fact that the meltwaters did not utilize the pre-existing valley to the west. Another factor is that this section of the channel was cut submarginal to ice within the pre-glacial valley.

At Hope the character of the channel alters significantly. Since it cuts across the volcanic ridge and is partly incised into the floor of the channel above Hope, it was probably formed at a later stage of deglaciation. It may even be referable to a separate glacial period, or to a re-advance of ice, but there is no evidence for these alternatives. Since it cuts across the ridge, the channel is best explained by a superimposition of englacial meltwaters onto the

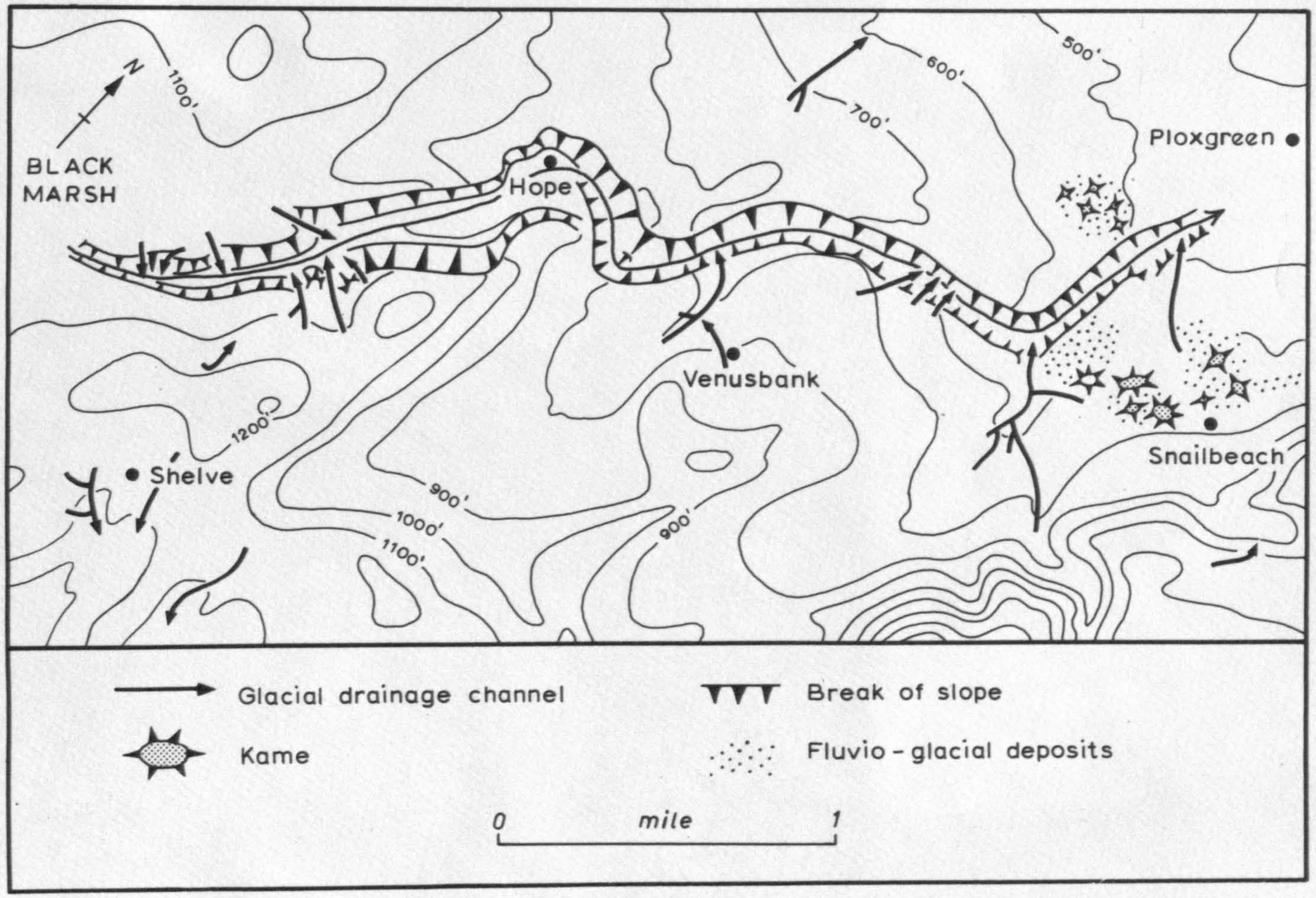


Fig. 5.12 Hope valley

underlying topography.

Beyond the prominent bend at Hope, the marked asymmetry of the channel suggests it was eroded either marginally or submarginally by meltwaters following the pre-existing depression, and flowing along the continuation of the volcanic ridge. The chutes which head into the southern side of the channel indicate a submarginal origin.

## THE STIPERSTONES

The Stiperstones ridge extends for some 7 miles from Pontesbury in the north-east to Black Rhadely Hill in the south-west, a trend conformable to the geological strike in the adjacent Shelve District. The ridge rises from a height of about 400 ft. in the north-east to a maximum of 1731 ft. at Mainstone Rock. The central part of the ridge lies between 1500 to 1700 ft., and several large tors,

> " ...stand out on the crest of the ridge at short intervals like rugged cyclopean ruins, some of the principal of which are about 50 to 60 ft. high, and about 120 to 130 ft. in width."[1]

The former presence of glacier ice enveloping much of the Stiperstones is indicated by several meltwater channels (fig. 5.13).

South of Pontesbury, on the western flanks of the Stiperstones, is a complex system of anastomosing channels, each some 10 to 15 ft.

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1. Murchison, R.I., 1839, p. 283.

deep and heading in a general northerly direction. Some are one-sided benches, while others are incised transversely across spurs. One channel heads parallel to the contours at about 600 ft. before turning abruptly to plunge directly down the hillside; in direct line with the upper section is another channel which isolates Callow Hill from the Stiperstones. Although eroded in bedrock, most channels have glacial drift in their floors, about 12 ft. being exposed in one north of Polesgate (SJ 392052). These factors strongly suggest that the channel system originated by the superimposition of englacial meltwaters onto the underlying topography.

To the south, a single large channel heads eastwards directly across the Stiperstones ridge and almost completely severing it. The channel is located at the back of an extensive flat at about 700 ft., behind which the Stiperstones rise steeply to about 1100 ft. The intake, at 775 ft., is at the head of a small pre-existing valley draining westwards, from whence the channel cuts across a minor spur at 800 ft. and heads eastwards towards Habberley, attaining a maximum depth of 75 ft. The outlet, at 625 ft., is cut partly into the flank of a fluvio-glacial terrace deposited against the Stiperstones.

It is probable that this channel was formed by the superimposition of an englacial stream across the underlying Stiperstones ridge. The meltwaters would tend to be concentrated at the break of slope at the back of the 700 ft. surface, a tendency noted elsewhere by Sissons,

> "... on steeper slopes one might expect the streams to move both laterally and downwards along the ice-ground contact towards the floors of cols or to areas of gentle slope."[1]

The erosion of the channel must also post-date the deposition of the fluvio-glacial terrace.

To the south, a prominent meltwater channel heads eastwards past Eastridge (SJ 383022) towards the Habberley Brook valley. The upper part of the channel, which has intakes at 1075 and 1100 ft., runs northwards down the Stiperstones towards Lordshill. Here, the channel turns abruptly towards the east occupying the floor of a broad valley. The channel attains a maximum depth of over 100 ft., and at Eastridge small chutes plunge down both sides. The outlet at 800 ft. is adjacent to Upper Vessons (SJ 387020), to the east of which is a broad, flat-topped fluvio-glacial deposit.

West of the headwaters of the Eastridge channel is another small channel which heads northwards parallel to the edge of the Stiperstones. At Lordshill Farm (SJ 379021) it makes an abrupt turn to plunge down the hillslopes, its outlet hanging well above the floor of the Snailbeach valley.

A number of small channels also plunge down the sides of Crowsnest Dingle (Pl. 22). These have intakes at up to 1400 ft., and several head for a short distance parallel to the valley side.

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1. Sissons, J.B., 1963, p. 108.

The outlets of most of these channels hang well above the valley floor, sometimes by as much as 100 ft.

The two chutes which plunge down the sides of the Eastridge channel strongly suggest that it was eroded subglacially. The channel itself shows a strong correlation with relief, and the subglacial meltwaters would tend to be concentrated in the floor of the col and in the pre-existing valley. The course taken by the channel at Lordshill Farm indicates it was eroded by marginal or submarginal meltwaters, which later turned to plunge directly beneath the ice. A similar origin will account for some of the channels on the sides of Crowsnest Dingle.

To the east of Tankerville Mine (SO 3559955) a steep-sided valley, over 100 ft. deep, has been cut into the western flanks of the Stiperstones. The contour pattern on the 1:25,000 topographic map gives no indication of the valley, but a continuous valley line may be traced cutting across the southern end of Buxton Hill, north-eastwards towards the Shelfield valley. At the head of this deep valley, on its southern side, three small channels, the intakes of which are at 1225 ft., plunge down the valley sides but do not grade to its floor. Lower down, two channels also head towards the valley from intakes at 1200 ft. in the Pennerley area; these, likewise, do not grade to the valley floor. It seems likely that all these channels were eroded subglacially.

South of the central high portion of the Stiperstones ridge is a prominent col at The Knolls (SO 363974), in the floor of which

Pl. 22 Meltwater channel on the flanks of Crowsnest Dingle

Pl. 23 Meltwater channel at The Knolls, Stiperstones

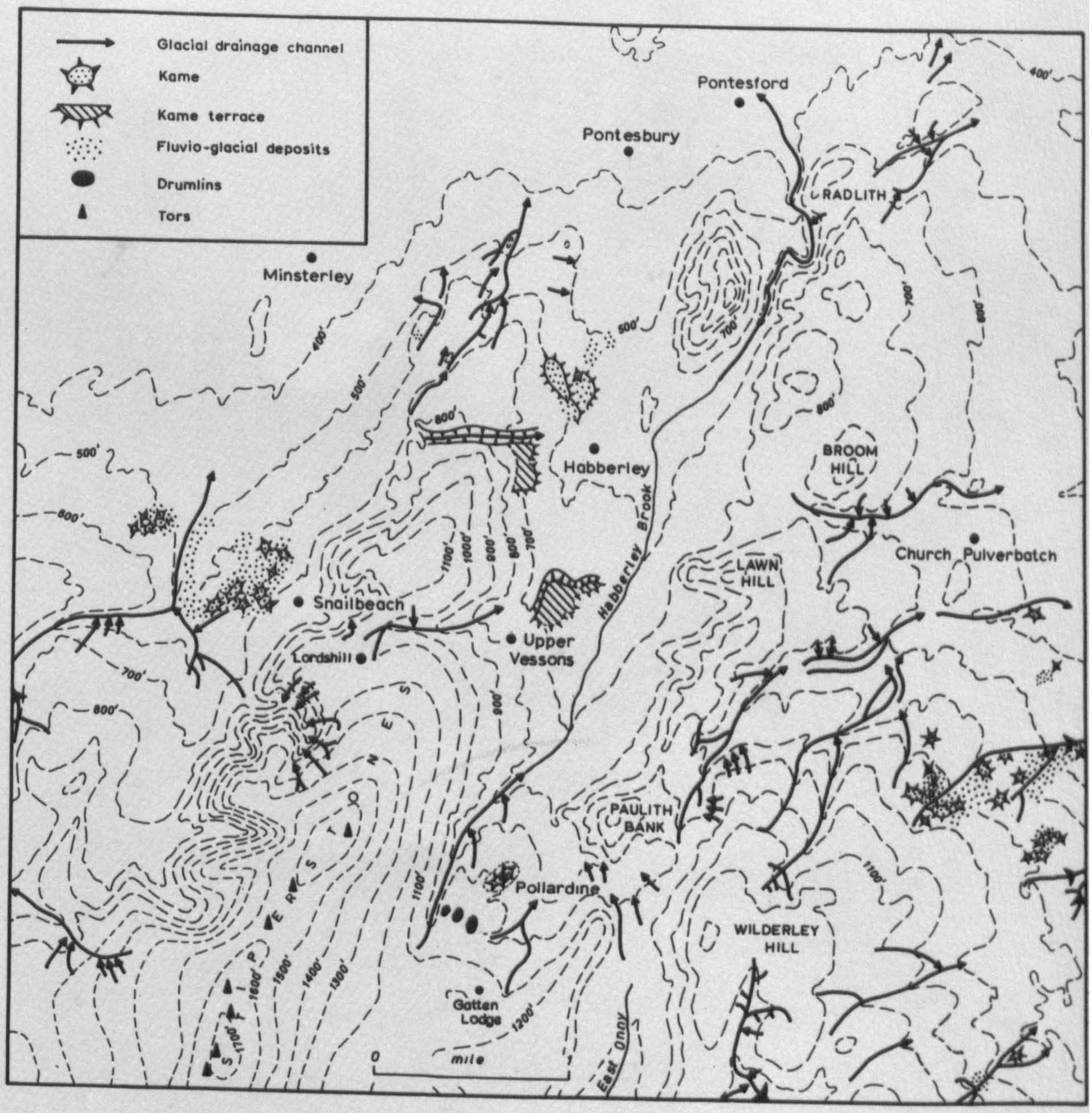


Fig. 5.13 Channels on the Stiperstones and Habberley valley area

is a system of meltwater channels, which plunge down the eastern flanks of the ridge from intakes at 1375 ft. (Pl.23). The intake of one hangs above the floor of another, and it isolates a rock knoll. These factors suggest that the system was eroded subglacially, beneath ice occupying the col.

## HABBERLEY VALLEY

There are several meltwater channels in the upper part of the Habberley valley. To the west of Pollardine (SJ 392002), the Habberley Brook flows in a deep, steep-sided gorge incised into the eastern flank of the Stiperstones (Pl.24). The intake of the gorge is at 1100 ft., and it deepens rapidly to a maximum of about 100 ft.; it opens out at 775 ft., just south of Brook Vessons farm. Several small chutes plunge down the eastern side of the gorge. North of Gatten Lodge (SJ 390993) two channels, head northwards along the valley floor and are confluent at The Sallies (SO 393997). Several small channels also head northwards from the upper East Onny valley into the Habberley valley. The intake of one is at 1025 ft., and it cuts across the watershed at the head of the East Onny valley.

East of Habberley a large channel, situated in the floor of a col at 725 ft. between Broom Hill and Lawn Hill, heads eastwards out of the valley. The intake is on a broad flat on the western side of Broom Hill, from whence the channel heads first south before turning eastwards. Several chutes plunge down the northern side of

Pl. 24 Meltwater channel in the upper Habbɒrley valley, adjacent to the Stiperstones

the channel and similar features occur on the southern side, the largest of which is over 20 ft. deep. The channel outlet is north of Church Pulverbatch (SJ 430029), and it is continued eastwards by a broad river valley.

The lower part of the Habberley Brook flows through a large gorge in the floor of a gap to the east of Pontesford and Earl's Hills (fig. 5.14.). The gorge occupies the floor of a somewhat broader and flatter valley that appears to have drained in part south-west, and in part north from a divide at Earl's Hill Farm (SJ 414052). The gorge heads generally northwards, but makes an abrupt turn to the east at the farm. Beyond this it is incised below a drift-covered flat at about 500 ft. Above the western side of the gorge is a broad col, Radlith, aligned NE - SW and the floor is at 525 ft. A small bench at 550 ft. is on the north side of the col, above the gorge. The gorge is steep-sided with slopes of about $32^{o}$, and at the eastward bend is over 75 ft. deep; beyond Radlith col it is only about 25 ft. deep and is considerably broader.

Northeast of Radlith col, and in alignment with it, is a complex series of meltwater channels (Pl.25). The intake of one is in the floor of the col at 500 ft., and another to the south is at 550 ft. These channels are confluent at Broompatch and head northeastwards towards Longden (SJ 442063). Several chutes plunge into both sides of the channel, and two, heading in opposite directions, isolate a bedrock knoll north of Lingcroft Pool (Pl.26).

Pl. 25 Channel leading away from Radlith col

Pl. 26 Knolls in the centre of channels at Radlith

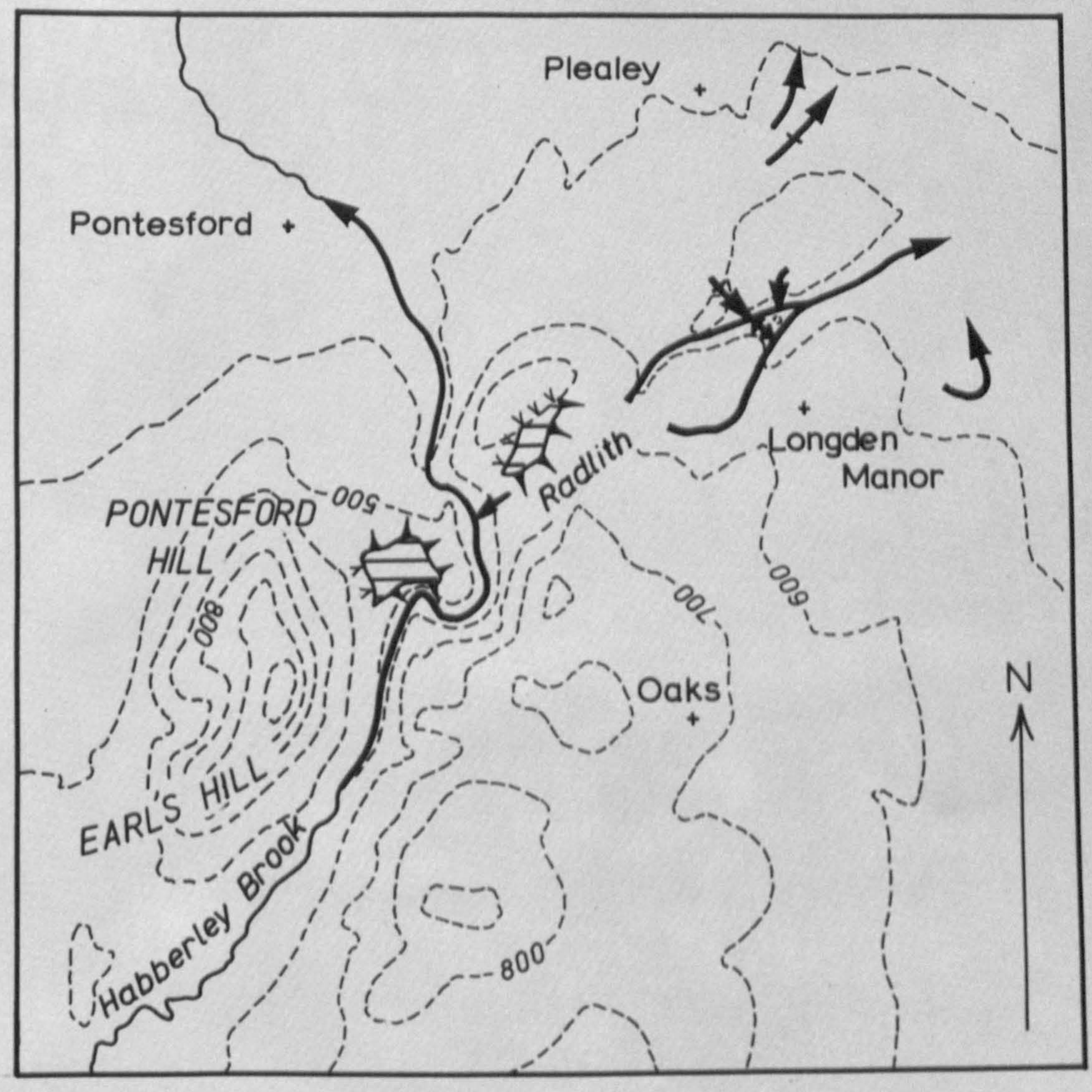


Fig. 5.14 Meltwater channels at Radlith col

At Plealey are two small channels, one of which incised through a spur and has a 'humped' long profile.

Since the meltwater channel between Lawn Hill and Broom Hill (fig. 5.13) cuts across the drainage divide on the east of the Habberley valley at its lowest point, it could be argued that it was formed as an overflow from a glacial lake impounded in the valley. Such an origin was proposed by Whitehead, who maintained that,

> "In the valley of the Habberley Brook ... a lake could hardly fail to be produced by ice, blocking its northern exit. Evidence of overflow from such a lake can probably be discerned in certain steep-sided channels in the high ground bordering the valley..."[1]

However, the morphology and course taken by this channel argues strongly against such an origin. If it was a glacial overflow it is difficult to see how, or why, it should first head southwards below Broom Hill before turning to the east. Also, the presence of ice above both sides of the channel at the time it was being cut is indicated by the several chutes on its sides. The meltwater channel is probably of subglacial origin, eroded beneath ice occupying the col between Lawn Hill and Broom Hill. The initial southward course

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1. Whitehead, T.H., in Pocock, R.W., Whitehead, T.H. and others, 1938, p. 203.

may have been eroded submarginally while part of Broom Hill had emerged as a nunatak. The meltwaters did not continue into the Habberley valley because the gradient of the ice surface, and thus the direction of the meltwater flow, was towards the east. Also, the ice was probably thicker in the valley than over the higher ground to the east.

Whitehead was also wrong in assuming that the all channels bordering the Habberley valley are glacial lake overflows. The channel on the Stiperstones at Eastridge supposedly formed an outlet for a lake at just over 1000 ft., and the Granham's Moor channel served a similar function when the level of the lake had dropped to 800 ft.[1] It has been shown, however, that the Eastridge channel flows into, and not out of, the Habberley valley; it does not cut completely across the ridge, and there is evidence that it was eroded submarginally. Likewise, the Granham's Moor channel flows into, and not out of, the valley. These channels cannot be regarded as glacial lake overflows, but were eroded subglacially, in association with a downwasting and stagnating ice mass.

With reference to the gorge east of Pontesford and Earl's Hills, it is obvious from the topography that the original course of the Habberley Brook lay to the west of those hills. It has been contended that,

> "... the diversion of drainage to the east of Earl's Hill was due to ice lying in the valley to the west."[2]

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1. Ibid.
2. Op. cit., p. 204.

This, however, would require ice, the surface of which must have been above 550 ft. to occur west of the hills, while the gap to the east, at a similar height, was ice-free. A more reasonable explanation is that the gorge was formed by the superimposition of an englacial stream onto the underlying topography. The prominent bend to the east was probably caused by the diversion of the stream when it encountered the pre-glacial divide at Earl's Hill Farm. The stream must have been flowing in the ice below this level, otherwise there seems no reason why it should not have been superimposed directly across the divide or flow northeastwards through Radlith col.

A subglacial origin also seems likely for the meltwater channels to the north-east of Radlith col. They form an anastomosing system, and it is probable that they were eroded by the superimposition of subglacial streams. This is suggested by the chutes which occur on both sides of the channel, and the 'humped' profile of that at Plealey.

## NORTHERN SLOPES OF THE LONG MYND

Several meltwater channels head northwards down the slopes of Cothercott Hill (fig. 5.15) the most impressive of which is Long Rough, 1,[1] (Pl.27). The intake of this channel is at 1325 ft., just north of the main watershed of the Long Mynd. A number of tributary channels, incised into a broad pre-existing valley-head, become

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1. Numbers refer to channels on fig. 5.15.

Pl. 27 Long Rough channel on the flank of Cothercott Hill

confluent below about 1200 ft., after which the channel deepens repidly to a maximum of about 50 ft. and heads directly down the hillslope, incised into the floor of a broad valley. To the west of Long Rough is a channel the intake of which, at 1200 ft., is in the floor of a shallow col. For a few yards the channel heads eastwards, parallel to the contours, and then turns abruptly to plunge down the hillside.

These two channels are tributary to a large channel, 2, at the foot of Cothercott Hill. Although morphologically continuous with that to the west of Long Rough, there is a marked deepening of the latter at about 900 ft., from whence it turns to head northeastwards. The channel has a broad, flat floor, mantled with glacial drift, and steep sides. It is asymmetrical in cross-profile, the southern, upslope, side being over 100 ft. high, compared with only 25 ft. on the north. At Cothercott the channel bifurcates around a bedrock ridge in its floor, and a narrow bench, at the same elevation as the surface of the ridge, is cut into the southern side of the channel. The channel outlet is at 700 ft. to the south of Castle Pulverbatch.

A number of small channels, 3, plunge down Cothercott Hill from a surface at 1000 ft., west of Cothercott Cottage (SJ 415010). Each is 10-15 ft. deep, and while some terminate part-way down the hillside, others grade to the floor of the drift-covered depression north of Cothercott Hill. A large meltwater channel, 4, the intake of which is at 1000 ft. in a col between Paulith Bank and Cothercott Hill, at the head of the East Onny valley, runs north-eastwards along the floor of the depression towards Castle Pulverbatch. For much

of its length the channel is confined to the northern side of the depression, and it is markedly asymmetrical in cross-profile. Several small chutes plunge down the northern side of the channel, one of which hangs above its floor.

Those meltwater channels plunging down the upper slopes of Cothercott Hill were probably eroded subglacially, by meltwaters finding their way beneath a stagnant ice-margin. The headwaters of Long Rough were concentrated into the broad valley head, and later into a pre-existing valley. The channel to the west probably originated close to the ice-margin, before the meltwaters turned abruptly to plunge beneath the ice.

The position and morphology of channel 2 suggests it was eroded either marginally or submarginally. Since it is morphologically continuous with that to the west of Long Rough meltwaters must have continued flowing into the upper part of the channel while it was being eroded, suggesting a submarginal origin. At Cothercott the channel bifurcates around a bedrock knoll in its floor, Similar features in Carlops, Peebleshire, are attributed to subglacial erosion,

> " ...the channels on either side of the 'islands' were formed contemporaneously,"[1]

a braided englacial stream becoming superimposed onto the underlying topography.[2] This may well have been the origin of the divided channels in this case. The ridge surface is at the same elevation

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1. Sissons, J.B., 1963, p. 106
2. Op. cit. p. 107-109.

as the narrow bench on the southern side of the channel, and it may well be that these are remnants of an earlier channel floor, into which the subglacial stream later became incised.

To the west, the channels, 3, which plunge down the hillside from the 1000 ft. surface are subglacial chutes, formed by meltwaters plunging into the depression north of Cothercott Hill.

Since the intake of channel 4 is in the floor of a col at the head of the upper East Onny valley, it might be thought that it is an overflow from a glacial lake. Whitehead maintained that clays in the upper East Onny valley,

> "... were deposited in an ice-dammed lake,"[1]

which may add support to this view. However, the clays in question are boulder clays.[2] No true lacustrine clays were found. The morphology and position of the channels described above suggest that the depression north of Cothercott Hill was occupied by a stagnating and downwasting ice-sheet, and it is likely that a similar state of affairs held in the East Onny valley. It seems probable that channel 4 was eroded subglacially, beneath stagnant ice. That this is true for the upper part of the channel is suggested by the fact that it is below the outlets of the chutes plunging down Cothercott Hill. The asymmetry of the channel suggests an ice-marginal or submarginal origin. The fact that chutes plunge down its upslope side supports the latter.

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1. Whitehead, T.H. in Pocock, R.W., Whitehead, T.H. & others, 1938, p. 197.

2. See p.

South of the watershed on Cothercott Hill is a large meltwater channel, 5, the intake of which is at 1325 ft. at the Thresholds (SO 411994). The channel occupies the floor of a broad valley (Pl. 28), forming one of the headwaters of Darnford Brook. Several chutes have been cut into both sides of the channel (Pl. 29), some of which hang above the channel floor.

Two smaller channels head eastwards off Stitt Hill into the channel 5. That to the north grades to the channel whilst that to the south grades to a bedrock flat on the channel side, possibly a remnant of the former valley floor.

To the south-east is channel 6, locally named 'Golden valley'. Its intake is at 1250 ft. in the floor of a col in the watershed; there is a sudden deepening in the long profile just below the intake and the channel makes a sharp turn to the north before heading northeastwards. The upper section of the channel is about 30 ft. deep and 20 ft. broad, with steep sides and a flat floor mantled with head and peat deposits. The channel opens out at about 1175 ft., and is continued by a broad valley.

That channel 5 is of subglacial origin is suggested by the occurrence of chutes on each of its sides. A similar origin probably accounts for the channels trending eastwards off Stitt Hill. This implies that the upper part of the Darnford Brook valley was occupied

Pl. 28 Meltwater channel at Thresholds

Pl. 29 Chutes on the flanks of the Thresholds channel

by a stagnant ice-mass during deglaciation. The Golden valley channel was eroded either subaerially or subglacially. Kendall was of the opinion that such channels,

> " ... which trench the main watershed of a country...",

were direct overflows from glacier-dammed lakes.[1] Such an origin was postulated for this channel by Lapworth and Watts who termed it a 'dry overflow'.[2] However, since the channel 5 and its tributaries were formed in association with a stagnant ice-mass in the upper Darnford valley, it seems unlikely that Golden valley is a glacial lake overflow. It strongly resembles a 'col gully', as recognised by Mannerfelt, and was possibly formed as the ice downwasted across the watershed. There is no direct evidence that it was eroded sub-glacially.

High Park Hollow and Hawkham Hollow are two broad valley on the north slopes of the Long Mynd and a number of meltwater channels occur within both. In the former, two small channels, 7, with intakes at 1300 ft. head obliquely down the valley side, one later turning abruptly to plunge to the floor. Further downvalley two other channels also head obliquely down the valley side, and grade to the broad drift-covered valley floor adjacent to Greenway Cottage (SO 441978). In Hawkham Hollow a small channel, 8, west of The Villa (SO 437977) heads for a short distance parallel to the valley

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1. Kendall, P.F., 1902, p. 481.
2. Lapworth, C. & W.W. Watts, 1910, p. 764.

side at about 1100 ft. and then turns to plunge to the floor.

Deadmansbatch, 9, is an impressive meltwater channel with a double-headed intake at 1300 ft. The channel heads directly down the hillside, deepening rapidly to a maximum of about 50 ft., and grading at its outlet to a broad valley southwest of Colliersley (SO 453979).

East of Deadmasnbatch is a similar, though less deep, meltwater channel, 10, with an intake adjacent to Blue Barn at 1100 ft. The channel heads directly down the hillslope and grades to the valley southwest of Colliersley. The latter is probably of pre-glaical origin, and two further channels, 11, are located on its northern side. The intake of one is at 950 ft. and it heads for a short distance parallel to the valley side, with a slope of about $1^{o}$, before turning abruptly to plunge to the floor. Adjacent to its intake, another channel plunges directly down the valley side. The intakes of these two channels occurs just below a prominent break of slope separating the relatively steep drift-free upper slopes of the Long Mynd from a gently sloping area on which are kame deposits.

The meltwater channels in High Park Hollow and Hawkham Hollow indicate that these two valleys were occupied, at least in part, by glacier ice. Most of the channels in these valleys were eroded subglacially. Those heading obliquely down the valley sides

are probably submarginal features, while those that plunge directly down the sides are subglacial chutes.

Deadmansbatch and the channel to the east were most probably formed by meltwaters flowing down the hillslopes directly beneath the ice. Channels 11 were possibly eroded at a slightly later phase of deglaciation. That which heads parallel to the valley side is an ice-marginal channel, and the meltwaters later turned to plunge directly beneath the ice. The other channel is a subglacial chute. The meltwaters were possibly directed into this valley along the break in slope at the foot of the Long Mynd, and the channels eroded contemporaneously with the deposition of the kames, near an ice-margin.

So far, most of the meltwater channels described have been situated on the upper slopes of the Long Mynd, above the limit of continuous drift. Below about 900 ft. the channels are cut mainly into drift, but often expose bedrock in their floors.

In the area of kame deposition adjacent to Underhill Hall, at about 950 ft., is the intake of a large meltwater channel, 12, which heads north-eastwards down the slopes of the Long Mynd (Pl.30). The channel is incised into glacial deposits throughout its length, and it grades at its outlet to an area of fluvio-glacial deposition north of Coppice House (SJ 445013). To the south-east, channel 13 has its intake in the area of kame deposition adjacent to Wildreck. Unlike the former, however, it exposes bedrock throughout much of its length, though glacial and fluvio-glacial deposits occur above

its sides. In the upper portion, a number of chutes plunge down its sides and several channels are tributary.

Semthcott Dingle, 14, is a channel, the intake of which is at 800 ft., adjacent to Smethcott Pool (SO 447996), from whence it heads north-eastwards (Pl.31). Small chutes occur on both sides of the channel, one of which is cut into a kame. At Parkhead, the channel turns abruptly towards the north, and its former course is continued by a dry channel heading towards Bank Farm (SO 463009).

Two small streams emerge from High Park Hollow and Hawkham Hollow onto a broad drift-covered area; they converge at 900 ft. and flow north-eastwards in a deep meltwater channel, 15, Bethcott Hollow. Sands and gravels are exposed in both valley sides, especially on the east. A small chute is on the western side of the channel. A series of boreholes across the lower part of the channel[1] reveal up to 21 ft. of glacial deposits. A system of tributary channels enters Betchcott Hollow at Branmill (SO 448992), beyond which they converge and head eastwards past Smethcott as a single, deep channel, 16. In the upper portion, bedrock is exposed in the sides and floor, and a small stream flows over a series of falls. Thereafter, however, the channel is cut entirely into glacial deposits, and cuts into an extensive series of kames and fluvio-glacial material around Smethcott. A number of chutes occur on the southern side of the channel, some of which hang above the floor. A series of boreholes across the channel showed 34 ft. of glacial deposits resting on a sloping bed-rock surface.

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1. For fuller details of the boreholes, see Appendix 1.

Pl. 30 Meltwater channel at Underhill Hall

Pl. 31 Smethcott dingle meltwater channel

At the mouth of the broad valley south-west of Colliersley a melt-water channel, 17, has been incised into the floor. This heads generally eastwards; it is flat-floored and steep-sided and cut into drift and into several kames. Several chutes occur on both sides of the channel.

Since the channels located on the upper slopes of the Long Mynd were cut subglacially, it is highly likely that those lower down have a similar origin. This is supported by the presence of chutes on the sides of most of them. Since they are cut into glacial drift, and sometimes into the sides of kames, they post-date the deposition of fluvio-glacial material. It is widely recognised that kame-and-kettle landscapes are indicative of thin, crevassed and stagnant ice[1], and it has been demonstrated that many of the meltwater channels associated with such landscapes originated subglacially. Galon holds the view that

> " ... in the time of deglaciation every longer halt of the inland ice, marked by marginal forms, produced an intensive erosional activity of subglacial waters along its edge."[2]

The system of meltwater channels on the lower slopes of the Long Mynd, especially those around Smethcott, probably were cut subglacially beneath the thin, crevassed and stagnant margin of an ice-sheet,

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1. Flint, R.F., 1942, p. 128. Niewiorowski, 1961, p. 427.
2. Galon, R., 1965, p. 19.

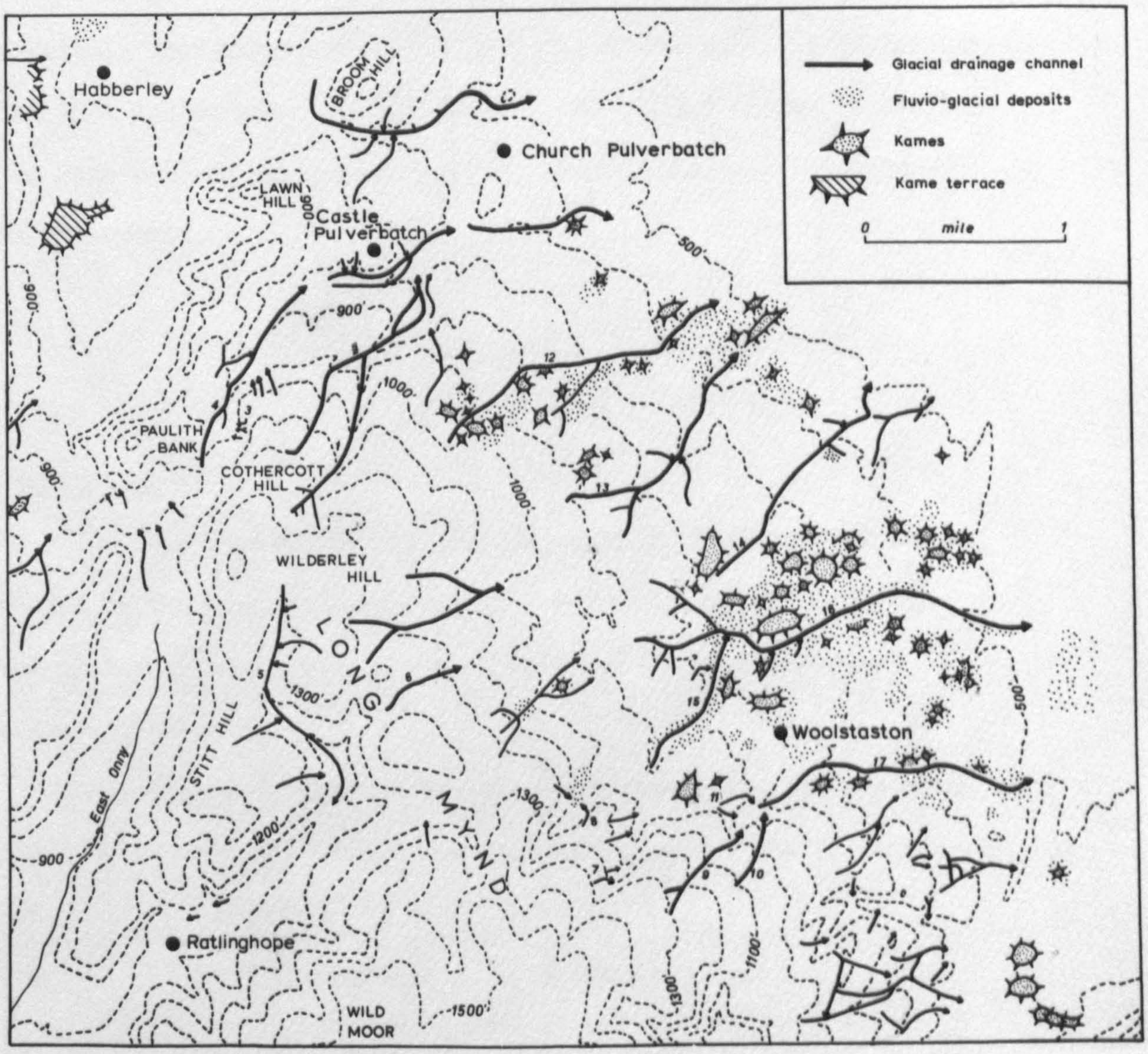


Fig. 5.15 Distribution of channels on the northern slopes of the Long Mynd

penecontemporaneous with the deposition of the fluvio-glacial deposits. They were probably formed at a later phase of deglaciation than that during which the channels on the higher slopes of the Long Mynd were eroded, consequent upon the downwasting and thinnihg of the ice-margin.

## VALE OF STRETTON

### Introduction

The Vale of Stretton is a funnel-shaped, fault-guided valley, bounded on the west by high-dipping Longmyndian rocks of the Long Mynd plateau, and on the east by a chain of hills composed of Uriconian volcanics.

Only passing reference has previously been made to the occurrence of meltwater channels within the Vale of Stretton. Watts mentioned that,

> "The western edge of the Long Mynd is fringed by lower, more or less isolated hills, cut off by what have the appearance of overflow channels, one of the most remarkable of which carries the 'Gravel Road', All Stretton."[1]

Similarly, Lapworth and Watts spoke of,

> "... dry marginal channels and cut-offs (North of Church Stretton); there are also

1. Watts, W.W., 1925, p. 331.

> gorge-like valleys (Marshbrook etc.,), and diverted stream courses (S. of Church Stretton - that of the ox-bow like valley west of Brockhurst is a conspicuous example.")[1]

Cobbold refered to,

> " ... the marginal channels, now dry, across many rocky spurs of the Long Mynd between the original lateral valleys,"[2]

and to the Marshbrook Gorge, which he believed to be cut across a col in the pre-glacial watershed,

> " ... at a height estimated at 800 feet."[3]

More recently, the Geological Survey and the Soil Survey have both turned their attention to the Church Stretton area. The former mapped many of the channels on the lower slopes of the Long Mynd,[4] and the latter, as well as indicating the Marshbrook Gorge, also show a channel between Hazler Hill and Helmeth Hill and between Helmeth Hill and Ragleth.[5]

Description of the Channels

The distribution of the meltwater channels is indicated on Fig. 5.16. They are well-developed and concentrated almost entirely

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1. Lapworth, C., & W.W. Watts, 1910, p. 764.
2. Cobbold, E.S., 1926, p. 233.
3. Op. cit. p. 232.
4. Sheet 166, Drift, (Church Stretton), Geological Survey of Great Britain, 1967.
5. D. Mackney & C.P. Burnham, 1966, fig. 7, p. 15.

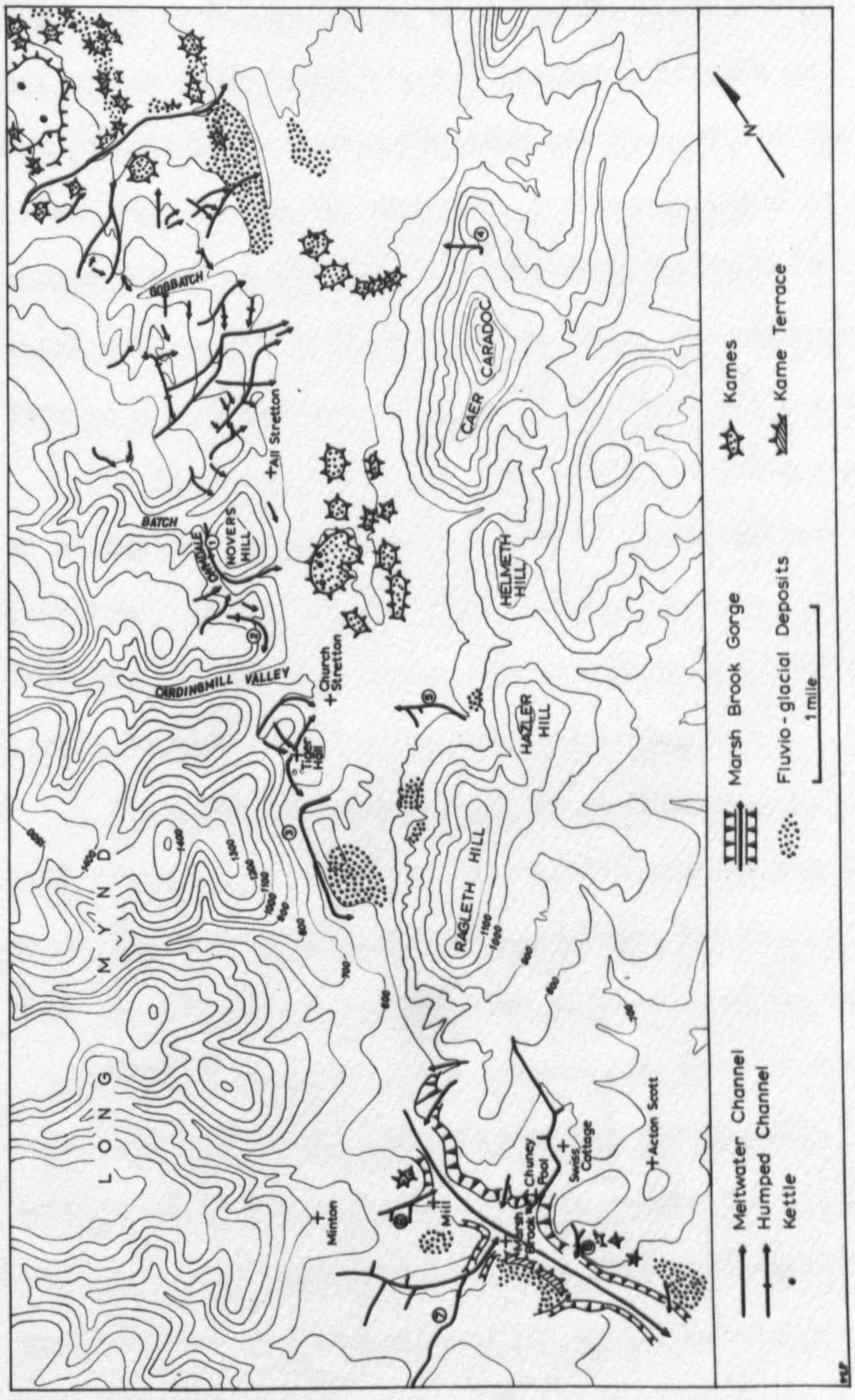


Fig. 5.16 Meltwater channels in the Vale of Stretton

on the lower slopes of the Long Mynd. A complex system occurs to the north of All Stretton, while to the south the principal features are the meltwater channel, 1,[1] isolating Nover's Hill from the Long Mynd, those above Church Stretton and that, 3, to the west of Brockhurst Castle. By contrast, the eastern side of the valley is almost devoid of channels. At the head of the Vale of Stretton a major valley, the Marshbrook Gorge, runs southward past Marshbrook, forming a link between the Vale of Stretton and the Onny valley.

To the north and north-west of All Stretton a complex system of meltwater channels has been formed on the eastern slopes of the Long Mynd, (fig. 5.17.). They occur at various levels between about 1050 ft., and in some instances long spurs are incised at several levels in an eastward-descending sequence.

At Jinlye (SO 450961) two small channels, 1,[2] with intakes at 1025 ft. and 1075 ft., head southwards towards the Batch valley, above the floor of which they hang about 150 ft. It is probable that the meltwaters ran into ice which filled the valley to this level.

East of Jinlye a large channel, 2, head southwards, from the upper Gogbatch valley towards Worsley (SO 455962). The channel intake, at 1000 ft., is some 30 ft., above the floor of Gogbatch (Pl.32). For the first 50 yd. it runs north-eastwards, cutting through the spur leading from Jinlye to Gogbatch and has a distinct 'hump' in its long profile. The channel then turns sharply towards

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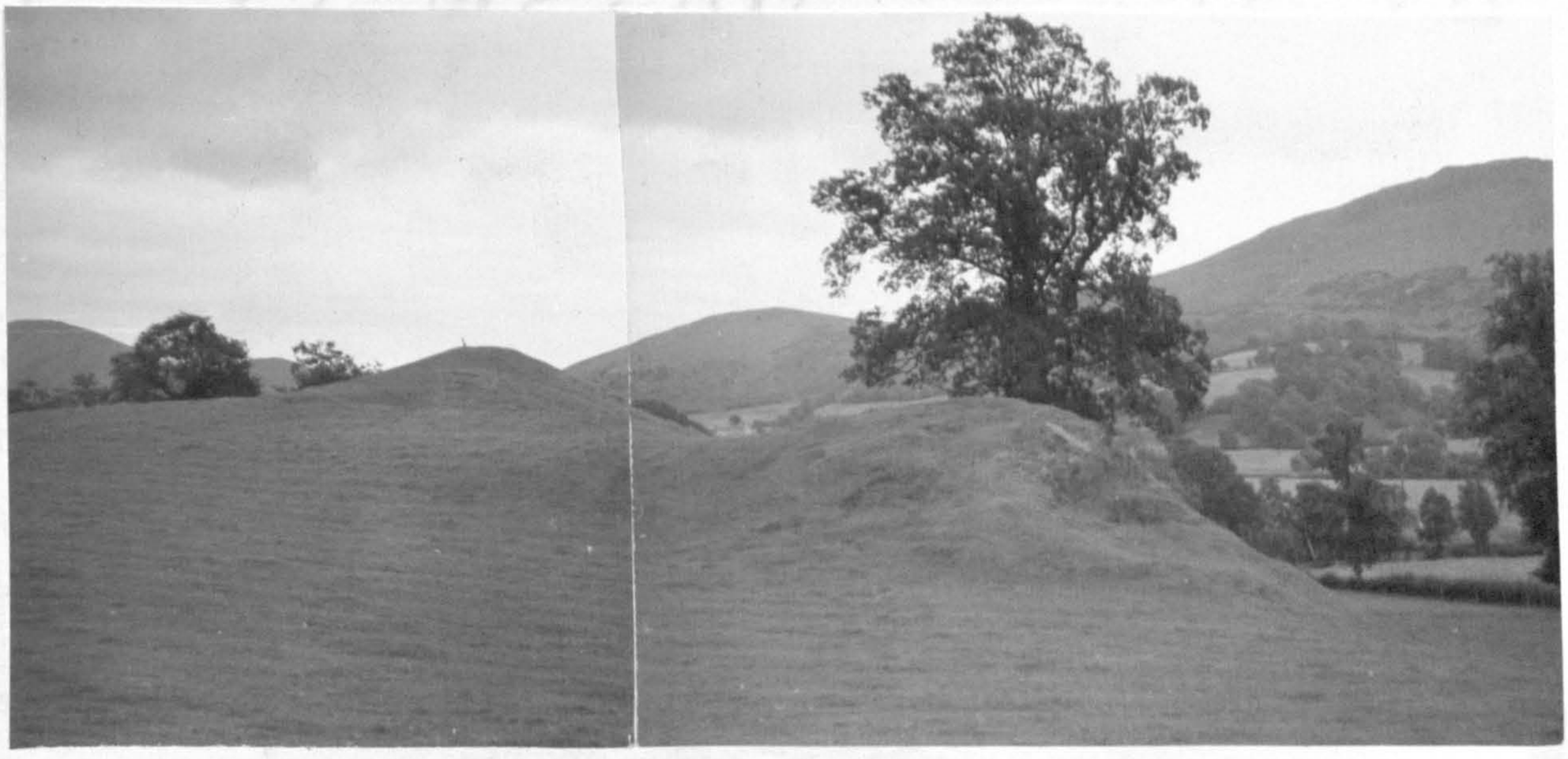
1. Numbers refer to channels on Fig. 5.16.
2. Numbers refer to channels on Fig. 5.17.

the south-southwest following the 900 ft contour for about 600 yd before descending into a tributary valley at Worsley. It is asymmetrical in cross-profile and its downslope margin is formed by a low bedrock ridge. Opposite the intake, and in direct alignment with it, the spur is cut across by two small channels, both of which hang above the floor of the main channel. Two channels head eastwards down the Long Mynd from the main channel, and the latter has two 'humps' in its long profile a few yards below the ponts from which the channels leave.

A complex system of channels and marginal benches, with a wide variety of form and position, is situated on the lower slopes of the Long Mynd to the east. Some head south, parallel to the contours, while others run obliquely down the hillslope. Although the system is an anastomosing one, the channels in it are not all accordant. The intakes of someare cut into the sides and hang above the floors, of adjacent channels. Many are not accordant at their outlets. Some, probably formed at a later stage of deglaciation, cut across others. Several are incised across spurs and have 'humped long profiles. In addition to the main channel system, a number of small chutes plunge down the sides of the larger channels, particularly on their upslope sides.

Rock knolls on the floor of the Vale of Stretton to the south of Dudgeley House are cut across by small channels (Pl. 36). These

Pl. 32 Channel cutting across a spur at Gogbatch, and the intake of the Worsely channel in the middle distance



Pl. 33 Knolls in the floor of the Vale of Stretton, cut across by meltwater channels

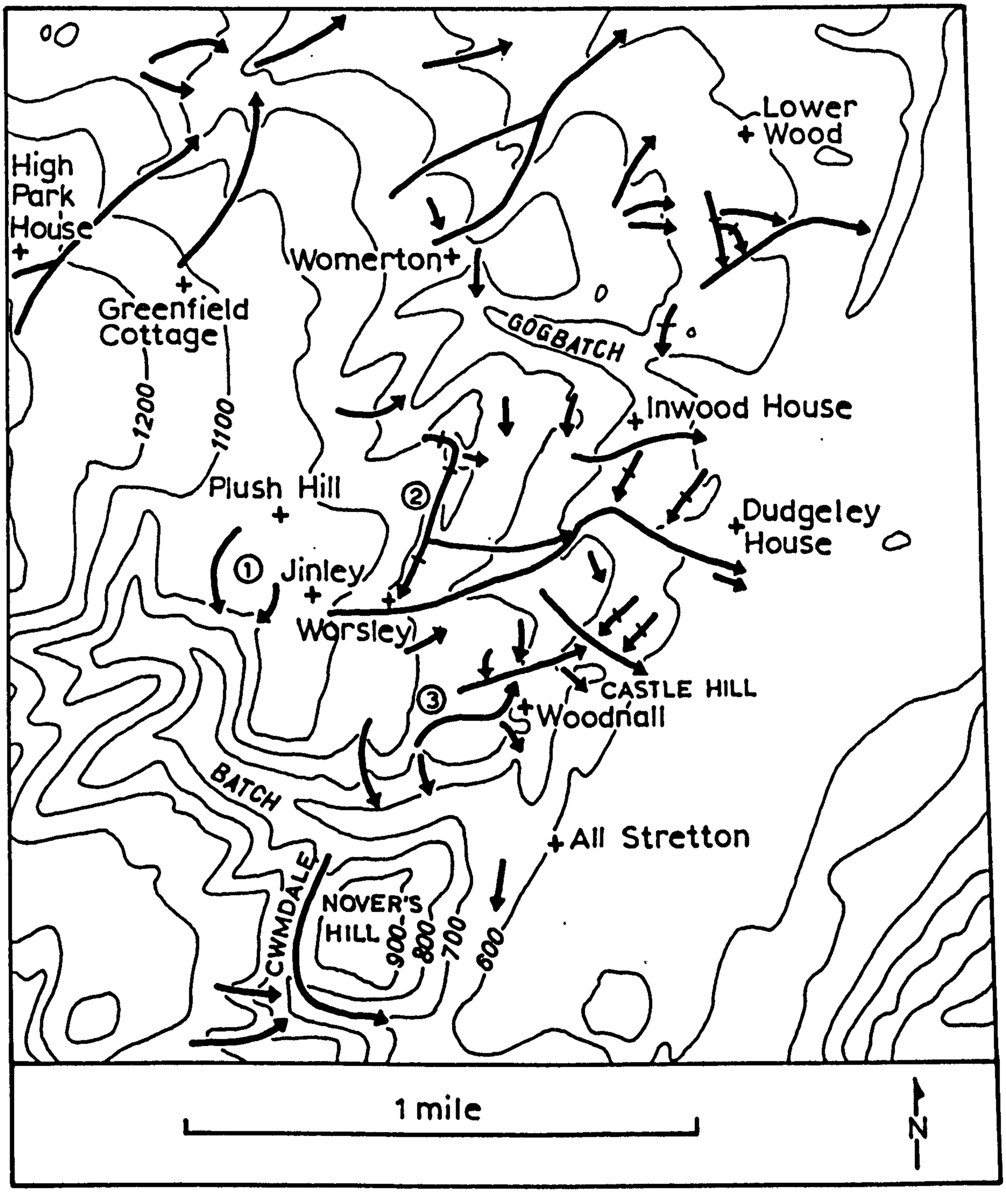



Fig. 5.17 Channels in the All Stretton area

are in direct line with two heading down the Long Mynd, with which they form an obvious continuation.

The channel, 3, eroded at the back of Synald's Coppice may be cited as an example of the complexity of the system. From its intake, which is about 100 ft., above the floor of the Batch valley, the channel heads towards Woodnall parallel to the contours, with a gradient of about $2^{o}$; small chutes plunge down the upslope side. At Woodnall the channel bifurcates; one branch heads eastwards at a similar gradient, cutting across a small spur at 800 ft., and its outlet grades to a bench on the side of Castle Hill; the other branch runs northwards with a slope of $8^{o}$ and is tributary to a channel running obliquely down the hillside. From the intake of channel, 3, a chute plunges down the side of the Batch valley and its outlet is 20 ft., above the valley floor.

A similar complex of channels occurs north of the Gogbatch valley, at Lower Wood. Again, some run parallel to the contours, while others plunge directly down the hillside; two channels are incised across a spur and have 'humped' profiles.

South of All Stretton a large meltwater channel, 1[1] Cwmdale, is over 100 ft. deep; it isolates Nover's Hill from the Long Mynd and appears to have been eroded at the break of slope between two surfaces at 950 ft., and 1000 ft. The intake is about 30 ft. above the floor of the Batch valley and for the first 100 yd. the floor rises at a slope of $5^{o}$; beyond this it falls at a similar gradient

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1. Numbers refer to channels on Fig. 5.16.

past the Church Stretton mineral works (SO 456947) into the main valley. (Pls.34 &35). Two shutes plunge down the upslope side from the 1000 ft. surface, and their outlets are about 25 ft. above the channel floor. (Pl.36 ). Below each outlet is a fan of coarse gravels. Up to 12 feet of similar material also covers part of the floor of the main channel.

To the south is a channel, 2, the intake of which is cut into the side and hangs above the floor of one of the chutes heading into the Cwmdale channel. For the first few yards it heads towards the south-east, cutting through a spur forming part of the 1000 ft. surface, at which point it has a 'humped' long profile. It then turns sharply and runs southwards, parallel to the edge of the Long Mynd, at a slope of about $5^{o}$. The cross-profile of this section is markedly asymmetrical, the upslope side being over 100 ft. high, whereas the downslope side is formed by a narrow rock ridge 25-30 ft. high. The channel outlet at 800 ft., is about 100 ft., above the floor of the Cardingmill valley.

Above Church Stretton a complex system of channels has been cut into a small erosion surface on a spur at 800 ft. Two small channels run southwards, parallel to each other at the back of the spur, towards Tiger Hall. The floor of that to the east is higher than that to the west; its intake is about 30 ft., above the floor of the Townbrook valley, and it has a 'humped' long profile. The intake of the western channel also hangs above the Townbrook valley, and a small chute plunges down the valley side from the intake. A narrow bench links

Pl. 34 Cwmdale channel, intake from the Batch valley

Pl. 35 Cwmdale channel. Note the prominent 'hump' at the farm

the intakes of the two channels; it is about 6 ft., above the floor of that to the west, but is accordant with that to the east. Beyond Tiger Hall a single channel heads southwards at the back of the 800 foot surface towards Cud Well (SO 448935), with a slope of about $1^{o}$. Beyond this point, which is at 750 ft., it plunges down the hillside at a slope of $15^{o}$. East of Tiger Hall a channel heads eastwards at about $8^{o}$ towards Church Stretton.

The lower part of the Townbrook valley appears to have been modified by glacial meltwaters. The upper portion, above the reservoir, is narrow, steep-sided and winding, with interlocking spurs. The lower section is broader, flat-floored and has no interlocking spurs and closely resembles a meltwater channel.

South of Church Stretton a large meltwater channel, 3, over 100 ft., deep, isolates Brockhurst Hill from the Long Mynd. Although a moraine is banked against the southern end of the hill the channel does not cut through drift, and bedrock was observed on its eastern side.

The general absence of glacial meltwater channels on the eastern side of the Vale of Stretton stands in marked contrast to the complex and well-developed systems on the lower slopes of the Long Mynd.

In the north, a small channel, 4, runs down the side of Caer Carodoc into the Vale of Stretton. The intake is at 1025 ft. on Little Carodoc, and the channel heads parallel to the contours for about 10 yds., before turning abruptly to plunge down the hillside.

The only other channel is that, 5, on the lower slopes of Hazler Hill. The intake, at 750ft., is adjacent to Plock's Coppice, from whence it heads towards Church Stretton. It is broad, flat-floored and steep-sided, and is cut into glacial deposits. Two small channels are tributary to its lower end.

At the head of the Vale of Stretton is a large meltwater channel, Marshbrook Gorge. This is cut into the floor of a broad gap between Ragleth Hill and Whittingslow Ridge, which probably carried pre-glacial drainage out of the Vale of Stretton.[1] Above its western margin is a drift-covered plateau at 675 ft, and two surfaces, at 725 ft. and 675 t., are on its eastern margin; it is followed by a branch of the Church Stretton Fault. The sides of the gorge are steep, with slopes up to 23° separated by sharp breaks of slope.

Close examination shows the gorge to be a complex feature. Several chutes, 6, plunge down its western side, and into a tributary valley at Queensbatch Mill (SO 441904). At Marshbrook Station, a system of channels, 7, heads into the gorge from the Marshbrook-Harderley gap. Small chutes also plunge down the eastern side of the gorge, the best example being the dry, double-headed channel, 8, west of Acton Scott (Pl.37). North of Swiss Cottage a meltwater channel is incised into the floor of a pre-existing valley, which is

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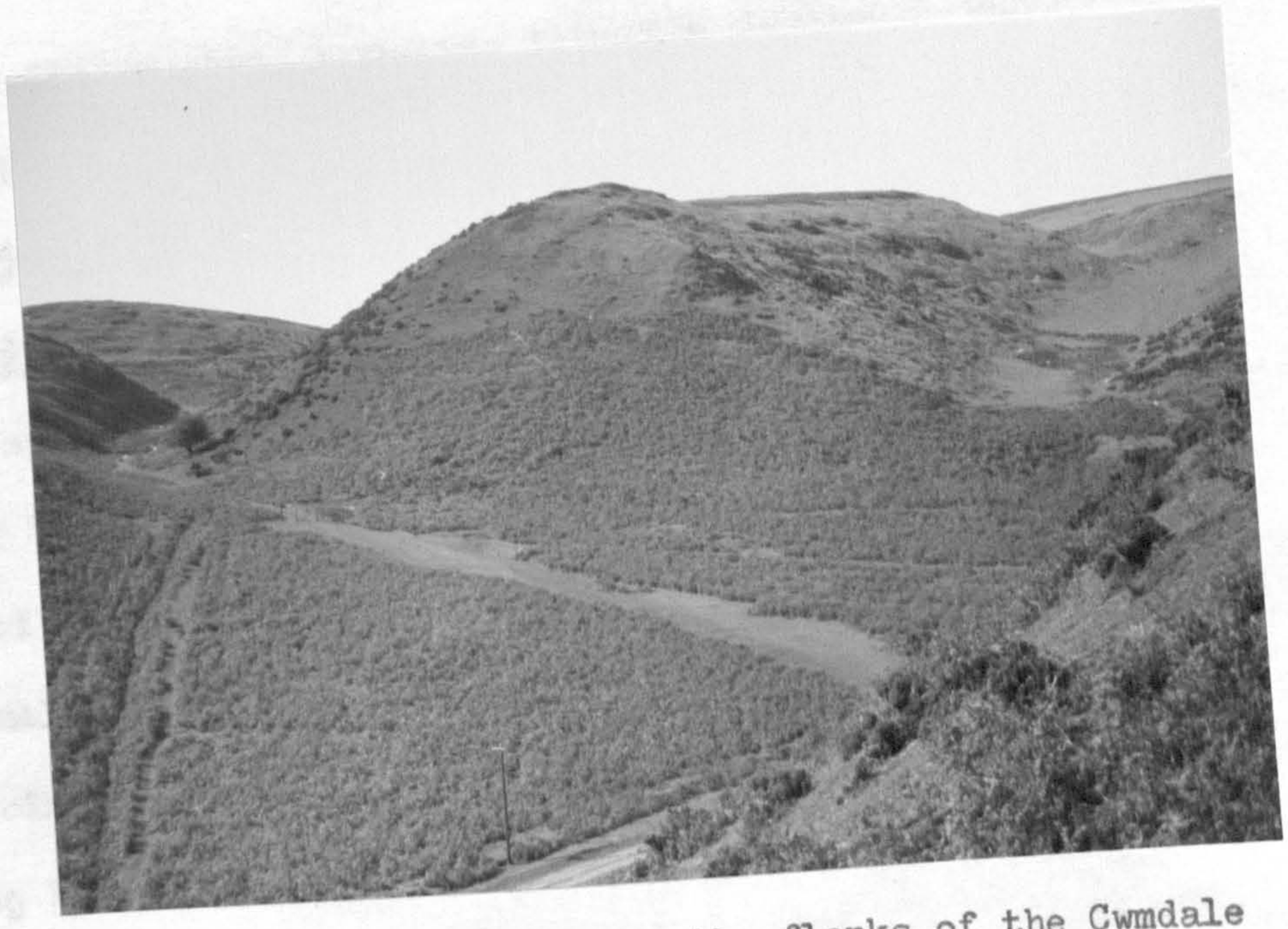
1. See p. 433
2. Cobbold, E.S., 1926, p. 233.



Pl. 36 Two chutes on the flanks of the Cwmdale channel

Pl. 37 Chutes on the flanks of the Marshbrook gorge

tributary to the Marshbrook Gorge; adjacent to Chuney Pool, a small channel heads parallel to the valley side for a few yards before turning to plunge to the floor.

Origin and Significance of the Channels

The form and position of the glacial meltwater channels within the Vale of Stretton give some indication of the extent of the ice, the slope of the ice surface and the character of deglaciation.

The meltwater channels above All Stretton were cut by meltwaters issuing from ice which had overridden the north-east shoulder of the Long Mynd to a height of about 1,100 ft. The occurrence of boulder clay in the lower part of Gogbatch and Batch valley[1] indicates that the eastern Long Mynd batches were occupied by ice, possibly lateral lobes of the Church Stretton glacier. Marginal and submarginal meltwaters evidently flowed southwards towards Church Stretton across the ice-filled valleys, cutting channels on the intervening spurs. Many of these have 'humped' long profiles, and were probably formed by the superimposition, across the spurs, of englacial meltwaters flowing under hydrostatic pressure, in the manner envisaged by Price.[2] In some cases there is a descending sequence of submarginal channels and benches, possible reflecting successive stages in the downwastage of the ice. Other channels head either obliquely or directly down the hillsides, often cutting across the submarginal channels and in some cases starting from them. These were probably

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1. See p. 162
2. Price, R.J., 1960 (a), p. 128.

formed at a late stage of deglaciation, when submarginal meltwaters were progressively finding their way directly into the ice, excavating subglacial channels. The small chutes which plunge down the sides of the larger channels were cut by meltwaters descending into stagnant ice within the channels at a fairly late stage of deglaciation. Other channels run down the sides of Batch and Gogbatch, indicating that the ice within these valleys became stagnant and meltwater drainage was no longer able to flow directly across them; instead it plunged beneath the ice from the intervening spurs. Although it is difficult to detect definite stages in deglaciation in this area, it appears that marginal and submarginal drainage was progressively replaced by meltwaters flowing subglacially as the ice stagnated and downwasted.

The morphology of channel 2, (fig. 5.17) is indicative of a complex origin. The alignment of the intake with the two small channels cutting across the spur and the fact that the intake hangs above the floor of Gogbatch, suggests an initial flow of meltwaters, on or within the ice, towards the east which became superimposed across the ridge (fig. 5.18). This would account for the 'humped' profile of the channel at its intake. The upper part of the Gogbatch must have been filled with ice at this stage. Subsequently, the meltwaters turned to flow southwestwards, parallel to the 900' contour, and the low gradient of the channel suggests it was formed in either a marginal or submarginal position. The latter is more probable, since the two channels which sever the downslope ridge and head directly

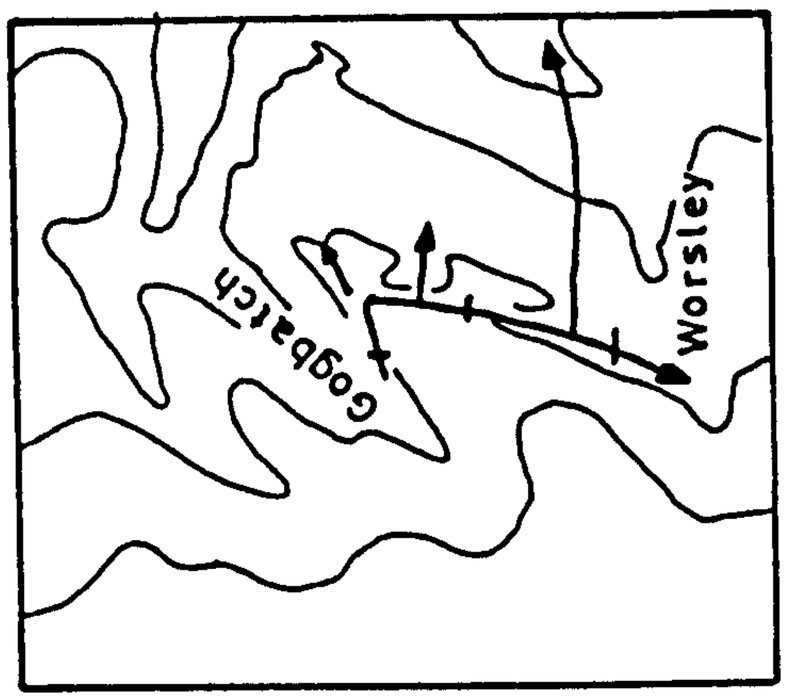


Fig. 5.18 Stages in the development of the Worsley channel

chutes plunge down its upslope margin. Beyond Woodnall, the branch which heads eastwards at a similar gradient was probably superimposed across the spur by englacial meltwaters, forming an original continuation of the upper section of the channel. Subsequently, it appears the meltwaters plunged directly into the ice, eroding that branch which heads northwards from Woodnall at a steeper gradient. At the channel intake, also, meltwaters subsequently plunged directly into ice occupying the lower Batch valley, eroding a chute on the valley side.

The large dry valley of Cwmdale which isolates Nover's Hill from the Long Mynd, was probably formed by meltwaters flowing southwards across the spur from the Batch valley. Several factors suggest it was formed submarginally. The floor has a 'humped' long profile which, although may be partly depositional, suggests erosion by englacial or subglacial meltwaters flowing under hydrostatic pressure. The ice margin at this time was probably at about 1000 ft. The two chutes which plunge down the upslope side of the channel also suggest the Cwmdale channel was cut beneath the ice. It is possible that the two chutes are periglacial features, but their intimate association with other meltwater channels, and the fact that the volume of material in the fans below their mouths is far less than would have been produced by the erosion of the channels, argues against this. The Cwmdale channel is one of the largest in the area, being about 100 ft. deep at the highest point in the valley floor and about 200 ft. deep at its lower end. It may not have been entirely excavated by

meltwater, but the southern portion may have originated as a normal stream tributary to the main valley. The later cutting through from the Batch valley, with some deepening of the channel, is ascribed to meltwater drainage.

Since the intake of channel 2 (fig.517) is cut into the side, and hangs above the floor, of one of the chutes descending into the Cwrumdale channel the erosion of the former must pre-date that of the latter. It probably originated by an englacial stream flowing off the 1000-foot surface becoming superimposed onto an underlying spur. Subsequently, the meltwaters turned to flow marginally or submarginally to ice occupying the Vale of Stretton, the relatively steep gradient of the channel suggesting the latter.

The two parallel channels at the back of the 800-foot surface above Church Stretton were probably formed by the superimposition of southward-flowing submarginal meltwaters across the spur. This would account for the fact that the intakes of both hang above the Townbrook valley, and for the 'humped' profile of that to the east. The meltwaters would tend to become concentrated at the break of slope at the back of the surface, and this would account for the greater depth of the channel to the west. The channel heading from Tiger Hall towards Cud Well was probably eroded by a continuation of the submarginal meltwaters. At a later phase of deglaciation, the meltwaters would find their way directly into the ice, eroding subglacial channels and chutes.

The meltwater channel isolating Brockhurst Hill from the Long

Mynd was probably formed by the superimposition of marginal or sub-marginal meltwaters, concentrated at the back of the 700 foot surface.

On the eastern side of the Vale of Stretton, the channel on Caer Carodoc was probably eroded by meltwaters flowing marginal to the ice for a short distance, before turning to plunge subglacially down the hillside. Similarly, the channel on the lower slopes of Hazler Hill was probably eroded by meltwaters flowing beneath the ice towards Church Stretton.

It has been established that the Marshbrook Gorge is situated in the floor of a gap at about 700 ft., between Ragleth Hill and Whittingslow Ridge, which probably carried pre-glacial drainage towards the south. Other valleys are tributary to the gap, the largest being that which heads north-eastwards from Cwm Head, and all have meltwater channels within them. The gorge therefore must have carried a considerable volume of meltwaters both out of the Vale of Stretton and from the surrounding areas towards the Onny valley.

The several channels which plunge down both sides of the gorge suggest that it was eroded subglacially, and that it was filled with dead-ice at a late stage of deglaciation. This is supported by the kame deposits which occur above both sides of the gorge. Such features are often found in association with thin and stagnant ice, and were possibly deposited in hollows and crevasses.

> "Glacial drainage channels indicate the position and condition of formerly ice-covered terrain."[1]

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1. Derbyshire, E., 1962, (a), p. 1111.

Previous work has been unanimous in the opinion that a glacial lake existed in the Vale of Stretton during deglaciation. Cobbold postulated that the lake formed a part of the much larger Lake Lapworth.

> "One arm of the lake stretched out up the Stretton valley, at least as far as Minton, and possibly to Plowden; the ice-front stood across from Acton Burnell to Church Pulverbatch, and the water having no access to the lower Iron-Bridge outlet had to seek another col in the surrounding hills. This was found in the neighbourhood of Marsh Brook, at a height estimated at 800 ft. The water overflowed at this point, and another gorge, much smaller than that at Iron-Bridge, was begun, this time across the Ordovician rocks of the Carodoc Range."[1]

A later advance of ice up the Vale of Stretton as far as Brockhurst forced more water through the gorge,

> " ... thoroughly establishing it. Along the edge of this glacier tongue the water had to escape southwards and cut the marginal channels ...,"[2]

Cobbold's hypothesis gained general acceptance: it is cited by Dwerryhouse and Miller,[3] and Wills indicates a pre-glacial watershed at Marshbrook, the 'overflow channel', and a lake in the Vale of

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1. Cobbold, E.S., 1926, p. 232.
2. Op. cit. p. 233.
3. Dwerryhouse, A.R., & A.A., Miller, 1930, p. 121.

Stretton co-existant with Lake Lapworth on a map showing the retreat phenomena of the Main Irish Sea glacier.[1] More recently, the Soil Survey made reference to a Marshbrook 'overflow channel'.[2]

There is no evidence, however, to support the concept of an ice-dammed lake of such proportions existing in the Vale of Stretton or for an 'overflow channel' at Marshbrook. The morphology and position of the meltwater channels indicates they were formed at or near the margin of a downwasting and stagnating ice-mass. It is difficult to ascribe a sub-aerial origin to any channel, and the majority were eroded submarginally or subglacially.

The initial stages of deglaciation were characterized by the erosion of a system of submarginal channels along the edge of the Long Mynd; this is represented in the north by channel 2 (fig. 5.17) above All Stretton and that behind Nover's Hill, and further south by channel 1, (fig. 5.16) those above Church Stretton and that isolating Brockhurst Hill from the Long Mynd.

At a later stage of deglaciation, submarginal drainage was replaced by meltwaters plunging directly into the ice, eroding sub-glacial channels and chutes. In some cases the chutes plunged down valley sides from the intakes of earlier-formed submarginal channels.

The glacial meltwaters were, in most cases, directed towards the southern end of the Vale of Stretton, towards Marshbrook. The Marshbrook gorge did not operate as an overflow channel, but was

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1. Wills, L.J., 1937, p. 92, Fig. 6.
2. Mackney, D. & C.P. Burnham, 1966, p. 6.

formed subglacially beneath a mass of stagnant ice in the pre-glacial gap between Ragleth Hill and Whittingslow Ridge, and it carried a considerable volume of meltwaters out of the Vale of Stretton towards the Onny valley.

The role of topography has been significant in the location and development of meltwater channels within the Vale of Stretton. The submarginal channels are consistently concentrated at the back of an erosion surface, reflecting the tendency for meltwaters to concentrate at concave breaks in slope. Many workers, including Bowen[1] and Sissons[2] have shown that meltwater channels are best-developed on long, gentle or moderate slopes, and the hillslopes above All Stretton provided favourable conditions for the development of a complex channel system.

Assuming that all the meltwater channels within the Vale of Stretton were formed, or at least utilized, during the same glacial episode, their distribution will give some idea of the extent of the ice; however,

> " ... it is quite another matter to think that in general the highest channels mark the maximum encroachment of the ice,"

since,

> "It is generally accepted that marginal drainage channels will only operate below the snow-line ...."[3]

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1. Bowen, D.Q., 1965, p. 372.
2. Sissons, J.B., 1963, p. 108.
3. Hollingworth, S.E., 1952, p. 107.

# Chapter 6

## MOVEMENT & EXTENT OF THE ICE SHEETS

### INTRODUCTION

Since there are no recognisable glacial cirques within this part of the Welsh Borderlands, it is safe to conclude that no significant masses of ice were nourished within the area. Clearly, the features of glacial erosion and deposition are referable to ice which invaded the area from outside, nourished in distant source regions.

A consideration of the colour, texture and lithology of the glacial tills and of included erratics indicates that they are derived from two distinct ice sheets, from separate source areas and moving in opposing directions. In the northeastern part of the area, and extending down the Vale of Stretton, the drifts are predominantly red-brown or brown in colour and commonly have a sandy texture. They include distinctive erratics such as granites and other igneous rocks from the Lake District and South-West Scotland, fragments of marine shells, and material derived from the Trias of the Shropshire and Cheshire lowland. These drifts were deposited by an ice sheet nourished in North-west England which moved across the Irish Sea basin, a lobe of which passed southward down the Cheshire-Shropshire basin.

By contrast, the eastern part of the area is dominated by drifts

which are brown or purple brown in colour, contain a high percentage of sedimentary rocks and are lacking in distinctive 'Northern' erratics. The presence of Plynlimon grits, and other erratics derived from the rocks of the Central Welsh Syncline, suggests the drifts were deposited by an ice-sheet which moved westwards from a source region in the Plynlimon area.

The glacial history of the district, and particularly the movement of the ice sheets which invaded it, cannot be separated from that of the Shropshire Plain to the north and the Welsh massif to the west. Consideration must, then, be given to the movement of ice in adjacent areas. (fig. 6.1).

To the north, boulders of felspathic larva and ash are scattered liberally over the Denbighshire Moors to a height of at least 1900 ft.[1] These felsite erratics are derived from the Arenig Mountains, and indicate a general movement of ice westwards across North Wales. The ice moved northeastwards towards the Vale of Clwyd and eastwards down the Dee and Ceiriog valleys.

On reaching the Welsh Borderland the ice sheet came into contact with 'Irish Sea' ice moving southwards down the Cheshire Basin. North of the Dee, the press of the Welsh ice and the obstruction caused by the Cefn-y-Fedw escarpment of Esculsham and Ruabon Mountains prevented the Irish Sea ice from invading the Welsh massif. Northern erratics and marine shells are found to heights of over 1200 ft. on this escarpment.[2] South of the Dee no such barrier exists and there

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1. These erratics have been traced in great detail by Macintosh. See: Macintosh D., 1873, 1874, 1879, 1882.
2. For a fuller description of these drifts see: Macintosh. D., 1882.

is some intermingling of the drifts. Welsh ice moving down the Dee valley debouched from the Vale of Llangollen and pressed across the Cheshire Basin. The Welsh ice, in pushing out eastwards, would be deflected southwards, and its drifts are intermingled with those of the Irish Sea ice.[1] The main line of confluence of the two ice streams extended southwards through Wrexham and Ellesmere towards Strewsbury.[2] That the drifts of the Welsh ice mass extended much further south is shown by the concentration of Arenig and other Welsh boulders in the neighbourhood of Brimingham and Bromsgrove.

In the Oswestry area, evidence from the distribution of erratics and the orientation of high-level glacial striae and moutoneed rock surfaces points to a general flow of Welsh ice towards the south-east, from the central Berwyn axis.[3] South of Llanfyllin, striae and roches moutonee indicate a movement of ice from the south-west. Wade, working in the Vale of Guilsfield area, refers to 'ridges of drift' (probably drumlins) on the crests of hills and on higher ground which are orientated towards the east-north-east.[4] These indicate a passage of ice from the Vyrnwy towards the Severn.

The Irish Sea ice never invaded the western hills in the Oswestry area, and the elevated outcrop of the Cefn-y-Fedw Sandstone again marks the western limit of this ice-sheet. Distinctive 'northern' erratics and marine shells have been found up to heights of over

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1. Wedd, C.B., et. al. 1929, Chap. 10.
2. Pocock, R.W. & D.W. Wray, 1925, Chap. 8.
3. Wedd, C.W., et. al. Chap. 10.
4. Wade. 1912, p. 211 & 214.

1,100 ft. in this area.[1] However, no trace of 'northern' material has been found in a southerly direction beyond Oswestry. Here, the drifts are entirely of Welsh character, probably deposited by ice streams moving down the Tanat and Vyrnwy valleys towards Shrewsbury.

Little work has so far been carried out on the movements of ice in the Welsh mountains to the west of the present research area, and the glaciation of the Plynlimon massif has still to be studied in detail. However, preliminary studies by Jones and Pugh indicated that an ice sheet was nourished in the Plynlimon area, from whence ice streams moved both westwards towards Cardigan Bay, and eastwards towards the Welsh Borderlands.[2]

The belt of high ground extending southwards from the Kerry Hills to the Wye gorge, forming the watershed between the rivers Teme, Lugg and Arrow to the east and the Ithon drainage to the west, is cut through by a number of high-level cols. Dwerryhouse and Miller[3] and, mo re recently, Luckman,[4] have shown that ice passed across these cols in an easterly direction, depositing glacial till and erratics and modifying them to various degrees by erosion. Erratics have been found on all save the highest summits along the watershed, and extend up to a maximum of 2000 ft. on Radnor Forest.[5] From the northern part of the watershed, the ice moved generally eastwards, channelled down the major valleys of the Clun, Teme and Lugg, and debouching onto the

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1. For a fuller description of these drifts see Nicholson, A.C., 1892.
2. Jones, O.T., & W.J. Pugh, 1935.
3. Dwerryhouse, A.R. & A.A. Miller, 1930
4. Luckman, B.H., 1966.
5. Ibid. p. 87.

lowland area between Wenlock Edge and Ludlow.

That the ice which crossed the watershed originated in the Plynlimon area has been demonstrated by members of the 'Committee on Erratic blocks of the British Isles'.[1] South of the Clun valley a group of erratics, notably Rhayader and Llandovery grits have been traced in an almost west-east line from Beguildy to Clunbury. Through they are most plentiful on top of the ridge of hills, they are by no means confined to high levels. The highest boulder, a grit from Rhayader is upon Black Hill at over 1,400 ft.

> "Professor Lapworth has examined a series of specimens, and describes them as Lower Llandovery grits and shales belonging to the Plinlimon group of central Wales. The hills from which they have been derived are all south of Bala, and situated almost due west from their present position."[2]

Smaller fragments lie in a pretty continuous stream right up to the Kerry Hills.

## MOVEMENT & EXTENT OF WELSH ICE

It is evident that Welsh ice entered the area from the west along two principal routes. One important ice stream must have passed down the line of the Severn valley, and this was supplemented

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1. Crosskey, H.W., 1882 & 1883.
2. Crosskey, H.W., 1882, p. 247.

by ice which crossed the high watershed between the Teme and Ithon drainage. (fig. 6.1).

That ice passed through the high-level cols in the watershed is demonstrated by the presence of till within them. The cols are all of pre-glacial origin, and have been modified to varying degrees by the passage of ice. The thick deposits of till in the floor of Black Gate col, at 1368 ft., the inclusion within it of several erratics closely resembling Llandovery grits, and the area of 'drumlinized' topography to the west, indicate that ice, which probably had its origin in the Plynlimon area, passed eastwards through the col into the head of the Mule valley. Similarly, ice must have passed through the col at Dolfor, and across the watershed at Crugynnau at a height of 1,570 ft., plastering drift on the eastern side of the upper Teme valley.

After crossing the watershed it would appear that the basal layers of the ice moved generally northeastwards, down the valleys draining towards the Mule-Gaebitra valley. There is some evidence that the upper layers of the ice were moving in a more easterly direction. The valley interfluves are dissected by a number of cols aligned west-east, many of which contain till deposits and have been modified by glacial erosion.

That ice passed down the upper Mule valley is indicated by the presence of till, and the oversteepened hillslopes south of Gilfach Bridge are probably the result of glacial erosion. The thick deposits of till and fluvio-glacial material also show that ice moved down the Miheli valley, and, again, the oversteepened valley sides are probably

ice eroded. The trough like col at 1,200 ft. in the interfluve separating the Miheli valley from that to the east is floored with till, suggesting that ice spilled eastwards through the col, into the next valley. This is confirmed by the presence of till within that valley. There are scattered deposits of till on the interfluves and in valley floors to the east, but perhaps because the valleys are narrower and aligned more north-south, transverse to a general eastward movement of ice, these deposits are only small and relatively insignificant. The broad col at Shepherd's Lodge, at 1,050 ft., has an oversteepened, almost vertical, south-eastern side and it is floored with extensive deposits of till which may be traced for some distance down the valley to the northeast. This suggests that ice passed through the col in a north-easterly direction, modifying it by erosion. Further east, at Argoed Lodge, the relatively broad, flat-floored col at 950 ft. is aligned west-east, and is floored by thick deposits of till which can be traced down the valley to the valley to the northeast. This again suggests that ice passed eastwards through the col, and that the basal layers were guided down the valley.

After moving north-eastwards down the valleys draining the Kerry Hills, the ice debouched onto the broad Mule-Caebitra depression, along which it was directed towards the east.

The position of the clusters of drumlins to the south of Glan Mule and to the east of Sarn is significant. The orientation of the first group indicates they were formed by ice which emerged onto the depression from the Miheli valley and that to the east, while the latter

The highest meltwater channels are at 1075 ft., at Jinlye, above All Stretton and on Little Carodoc. Further south, the submarginal channel behind Nover's Hill was initiated at 1000 ft., chutes head into it from the 1000 ft., surface, and channel 2 cuts across a spur at the same height. At Chutch Stretton the highest channels are at 800 ft, while the Brockhurst channel was initiated at 700 ft. Thus, it may be deduced that the ice extended up to at least 1100 ft., in the north of the Vale of Stretton, and the surface sloped down towards the south, to a height of at least 700 ft., at Brockhurst. At the head of the Vale of Stretton, ice must have occupied the Marshbrook gap up to at least 700 ft. This compares favourably with the evidence of the distribution of drift and erratics.

group was formed by ice moving slightly no rth of east towards Church Stoke. There is no conclusive evidence, however, as to the origin of the drumlins and the precise mechanism of their formation remains in doubt. There is little evidence of the internal structure or composition of the drumlins, though some are composed almost entirely of rock with only a veneer of till, and others have rock cores. Not all are erosional, since the plastering of smaller drumlins on the sides of larger ones, giving a two-tiered effect, must be a depositional feature. The similarity of these drumlihs to the streamlined bedrock forms which characterize the hills to the north suggests that they must be, in part, of e rosional origin. Hollingworth expressed the view that,

> "The existence of considerable hills of rock that have been moulded into typical drumlin shapes that are parallel to, and frequently indistinguishable from, the associated true drift drumlins, necessitates a prolonged and powerful ice-movement in a constant direction. Conditions under which erosion and deposition went on almost simultaneously are suggested by this association ...."[1]

It would appear that the drumlins in the Mule-Caebitra depression were formed beneath an active and moving ice mass by some process which combines a moulding action more or less associated with deposition. The fact that in both cases the drumlins are located in a position where the ice could spread out after being constricted in relatively narrow valleys may be significant in their formation.

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1. Hollingworth, S.E., 1931, p. 341.

The orientation of the drumlins south of Glan Mule indicates the movement of an ice stream towards the north-east. On reaching the broad Mule valley, this ice would meet another stream moving eastwards from Kerry, and the two would probably continue in a resultant direction slightly north of east towards the Caebitra.

Thick deposits of till in the low gap west of the hill at Pen-y-Gelli suggest the passage of ice through it. This ice also probably deposited the tough basal till at Goetre at the entrance to the Mule gorge.

The orientation of the numerous rock drumlins on the lower slopes of Caebitra Uchaf, above Llwyncowrid, indicates erosion by ice moving north-eastwards towards the Caebitra valley. The elongate bedrock ridge to the south of Sarn, which has the form of a tapered spur, was probably fashioned at the junction between two ice-streams, one entering from the Mule valley, and the other moving northeastwards down the valley leading from the Shepherd's Lodge col. The Cefn-y-Coed hills are dominated by streamlined ridges, or 'rockdrumlins', all of which trend towards the north-east. Although there are few distinctive erratics within the drifts of this area, it is evident that an ice-stream entering the region from Central Wales down the axis of the Severn valley would become confluent with the ice moving north-eastwards from the watershed of the Kerry Hills. The combined ice streams would then pass north-eastwards towards Montgomery.
The rock-drumlins are best developed on the lower hill areas, adjacent to the Severn valley, presumably because it was here that the ice was

thickest. The ice would be thinner over the higher hills south of Montgomery, and presumably its power of erosion would be reduced. However, even the highest hills show some evidence of glacial erosion, and ice must have covered the Cefn-y-Coed hill mass. The ice passed transversely across the Mule gorge, and a suggestion that the gorge was cut subsequent to the advance of ice is the truncation of several rock-drumlins by the steep sides of the gorge. Several rock-drumlins were also moulded by the ice in the floor of the Severn valley, and the valley sides show some evidence of oversteepening.

Beyond Montgomery, the lack of distinctive indicator erratics in the drifts makes difficult the tracing of the movement of the Welsh ice, and one has to rely on the colour and nature of the tills and the orientation of the ice moulded landforms. The latter will only indicate the direction of movement of the basal layers of the ice, and that of the upper ice layers must be inferred.

In the Montgomery - Church Stoke gap several rock ridges, masked with a brown clay till with erratics of local Silurian shales and grits, are aligned in a NE-SW direction. It seems likely that these rock ridges were fashioned, at least in part, by ice moving north-eastwards from the Cefn-y-Coed hills.

The area between Marrington Dingle and the steep western edge of the Corndon - Shelve massif is characterised by numerous elongated rock ridges, of which the Ridge is the largest. Several roches moutonee occur on the crest of the Ridge. The rock ridges are orientated almost north-south and continue beyond Rorrington into the Marton valley.

It would seem that these were eroded by ice which had been deflected northwards as it pressed against the steep western margin of the massif. It is evident that an ice stream also entered the Rea Brook-Camlad valley from the Severn.

The orientation of the ice-moulded rock ridges north of Montgomery indicates that the ice passing down the Vale of Powis was divided into two streams as it abutted against the Long Mountain: one stream continued northwards towards Welshpool and the Breidden Hills, and the other passed down the Rea Brook - Camlad valley towards Asterley. Elongated rock ridges can be traced almost continuously from Montgomery Station along the south side of the river Camlad towards Rorrington. They can also be traced east of Forden, extending up the Severn valley.

The glacial till in the Marton valley is wholly of Welsh origin. There is a lack of distinctive indicator erratics and a preponderance of local shales, grits and sandstones. However, the orientation of the numerous ice-moulded ridges superimposed on the platform on the northern side of the valley demonstrates that ice moved north-eastwards. Around Minsterley is a group of drumlinoid mounds, orientated NNE-SSW. It seems likely that these were deposited by ice, moving down the Hope valley from Corndon - Shelve massif, into the Rea-Brook valley.

The orientation of the elongated rock ridges on the western side of the Severn valley around Welshpool also indicates that an ice stream was directed down that valley towards the north. Since the Middletown gap is covered with Welsh drift it is probable that the Severn ice mass bifurcated against the Breidden hills and that one

stream was deflected into the gap.

In the northern part of the district the movement of Welsh ice from the Plynlimon area becomes increasingly affected by ice streams entering the area from further north, from the Vyrnwy valley and the area west of Oswestry, which were moving towards the south-east. Also, in moving into the Shropshire - Cheshire Plain, the Welsh ice was affected by the ice moving southwards from the Irish Sea basin.

Glacial striae in the channel of the river Severn at Belan Bank, at the northern edge of Breidden hill, indicate an eastward movement of ice in this area.[1] The distribution of distinctive erratics of Breidden dolerite between Minsterley and Asterley, and around Westbury, and of several specimens of Carboniferous Limestone from Llanymynech, indicate a movement of ice from northwest-southeast. Several other erratics of Carboniferous Limestone have been noted at and near Cruckton and south of Arscott; while an erratic of Alberbury Breccia in a well at Park Farm, near Yockleton must have also come from a site somewhere between north and northwest.[2] This movement of Welsh ice towards the southeast probably resulted, as in the Oswestry area, from the interplay of the Welsh ice and Irish Sea ice.

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1. C.B. Wedd, et.al., 1929, p. 151.
2. Pockock, R.W., et.al., 1938, p. 191 & 192.

In tracing the movements of ice over the Cordon - Shelve hill mass, and over the Longmyndian country to the east, greater use can be made of the distribution of distinctive indicator erratics (fig. 6.2) and less reliance is placed on the orientation of ice-moulded landforms. The drifts covering the area between Corndon hill and the Long Mynd are entirely of 'local' or Welsh origin, and were evidently deposited by ice moving into the area from the south and west. The deposits are located chiefly in the floors of the principal valleys, suggesting that these were the lines of maximum ice penetration. However, scattered erratics over the hillsides testify to the former presence of ice at higher elevations, and aid in elucidating its movement which, at higher levels, was often transverse to the valley lines.

It seems likely that the ice stream which was moving east- north - east down the Caebitra valley should continue eastwards along the broad Camlad - Onny depression towards Lydham, as well as pressing north-eastwards against the western edge of the Corndon Hill massif. This eastward movement is suggested by the fact that the till covering the southern side of the Caebitra valley can be traced continuously onto the south side of the Camlad valley. The drifts include erratics of Silurian shales, grits and flags, derived locally or from the west. The till on the opposite side of the Camlad valley is similar in character, but contains a proportion of lava and tuff probably derived from local Ordovician outcrops.

On the lower, southern slopes of Heath Mynd and on Norbury Hill, up to over 1325 ft. are several erratics of volcanic tuff, rhyolite

and other dark green, medium-grained volcanic rock (possibly dolerite). The source of these was most probably the Ordovician volcanic outcrops to the west, and they provide strong evidence of an eastward movement of ice up the Camlad valley. Whitehead found erratics of Stiperstones Quartzite and boulders of dolerite, similar to that of Disgwylfa Hill, on Oakley Mynd.[1] This indicates a southeasterly movement of ice from the Camlad past Lyndham towards the southern end of the Long Mynd.

The distribution of erratics indicates that there was also a movement of ice north-eastwards into the Corndon Hill-Shelve massif. While most of the erratics in the Shelve area are of igneous rocks, the source of which is difficult to locate, several specimens of a distinctive picrite occur which could only have been derived from the southwest, at Cefn. This suggests that an ice stream moved into the Shelve country along the lowland area between Corndon Hill and Cefn Gunthley, depositing drift in the valley floors.

There is also evidence that ice entered the Shelve area from the east, around both sides of Corndon Hill. It was suggested earlier that Corndon Hill showed evidence of 'crag-and-tail' features on a large scale. The western face was exposed to the abrading and smoothing action of ice, while the eastern, or leeward, side was a zone of plucking. Also, the distribution of drift indicates a 'tail' on the eastern side of Corndon extending across Hyssington Marsh. Neighbouring hills such as Roundton and Todleth show similarly contrasted

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1. Greig, D.C., et. al., 1968, p. 291.

topography on the two sides. It seems likely that some of the ice that was pressing north-eastwards against the western flank of Corndon managed to escape eastwards across low cols and moved down the low-land areas to the north and south of the hill towards the Shelve district.

In the East Onny valley distinctive Pentamerus sandstone erratics are found in an almost continuous line from the obscured outcrop along the southern edge of the Longmyndian rocks northwards towards Bridges. There are also several erratics of limestone with brachiopod fossils on the southern slopes of Norbury Hill, and occasional specimens of crinoidal limestone in the East Onny and Criftin Brook valleys. These erratics indicate a movement of ice from the south, up the western edge of the Long Mynd. The limestones are probably of Wenlock age, derived from outcrops to the south, and were introduced through the two broad gaps on either side of Oakeley Mynd. This ice was confluent with that advancing eastwards past Lydham, and the combined streams moved up the East Onny valley and around the southern edge of the Long Mynd towards Marshbrook (fig. 6.1).

In the Clun Forest area the distribution of Plynlimon erratics indicates a general eastward movement of ice at a relatively high level. Dwerryhouse and Miller state that ice moved eastwards down the Clun valley towards Clunbury[1], and the Soil Survey show that this stream

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1. Dwerryhouse, A.R., & A.A. Miller, 1930, p. 100.

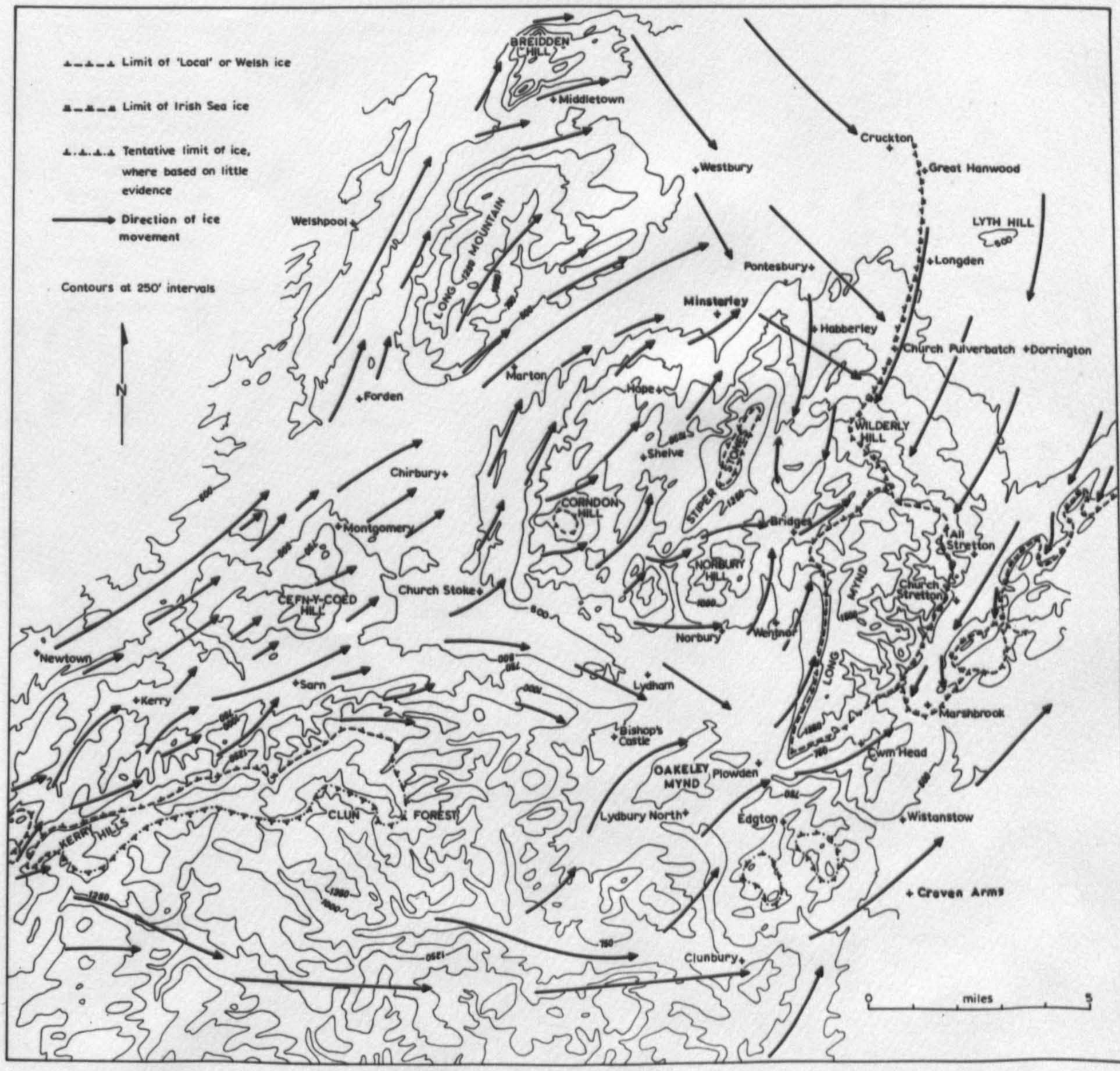


Fig. 6.1 General direction of ice movement

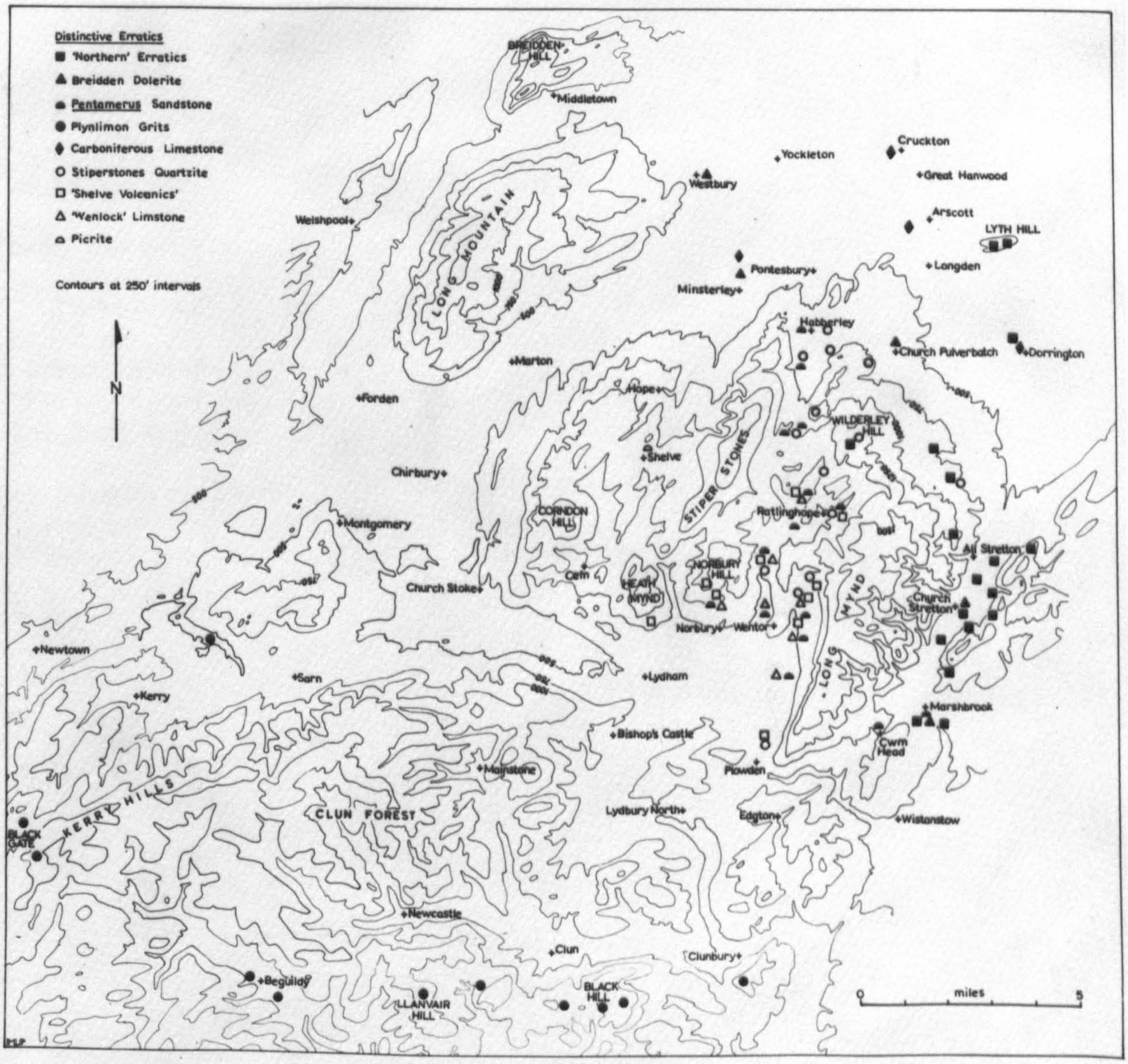


Fig. 6.2 Distribution of erratics

continued towards Craven Arms into lower Ape Dale,

> " ... as Wenlock Edge and View Edge prevented further penetration to the east."[1]

It would appear that ice from the Clun also moved northeastwards through the gaps on either side of Oakeley Mynd.

There are also numerous erratics of rhyolite, ashes, dolerites and other igneous rocks, together with some Stiperstones Quartzites, in the East Onny and Criftin Brook val leys. The rhyolites and other igneous rocks were probably derived from the Corndon-Shelve district, and, together with the quartzites, imply an ice movement from the west and southwest. One stream probably passed through the gap between Norbury Hill and the Stiperstones, while the other moved into the area from the Camlad valley.

It is possible that ice moved over the watershed between the East Onny and Habberley valleys. Although the watershed itself is devoid of drift, a group of drumlinoid features near Gatten Lodge are orientated north-northeast to south-southwest, with their steeper ends facing SSW. They appear to have been deposited by ice moving NNE, against the flank of the Stiperstones. Several erratics of Pentamerus sandstone in the upper Habberley valley could also have been derived from the south.

In the Habberley valley the prevalent type of drift is a brown or yellow-brown till of Welsh origin. The erratic distribution both in the valley and over the area immediately to the east points to a south

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1. Mackney, D., & C.P. Burnham, 1967, p. 17.

and southeastward movement of ice. Numerous Stiperstones Quartzite erratics occur on the eastern side of the valley, for instance east of Pollardine and Habberley, and local Longmyndian boulders are confined to the eastern side, close to the outcrop. This suggests a movement of ice from the west or northwest, across the Stiperstones ridge. Several erratics of Pentamerus sandstone occur on either side of the valley, as at Pollardine, Marsley and near Habberley. These could have been derived from the outcrop on the southern side of the Long Mynd - Shelve district, via the East Onny valley, or from the outcrop on its northwestern margin, brodering the Rea Brook - Camlad valley. They may have been carried by ice moving eastwards across the Stiperstones, or moving south up the Habberley valley.

Drifts covering the lower slopes of the chain of hills bordering the eastern side of the Habberley valley are dominantly brown till of Welsh origin. Several erratics of Stiperstones Quartzites found in the drifts, as at Oaks, east of Castle Pulverbatch and on Wilderley Hill, again point to an ice movement from the northwest. This is supported by the occurrence of dolerite boulders at and near Oaks from Pontesford and Earl's Hills. The large boulder of green dolerite at Church Pulverbatch has evidently been carried from the Breidden Hills to the northwest. The direction of movement of Welsh ice in this area continues that established in the area between the Breiddens, Pontesbury and Great Hanwood.

Drift in the Onny basin west of the Long Mynd may be traced through the Plowden gap, up the Horderley - Marshbrook valley towards

the Vale of Stretton. The till is brown, yellowish, or purple in colour and contains a substantial proportion of local erratics, together with some quartzite boulders. The latter are probably derived from the Stiperstones and introduced by an ice lobe which advanced into the Onny valley from the west. Boulder clay is continuous between Minton and Little Stretton, though locally it is thin. It seems likely, therefore, that ice moved into the Plowden gap from the Onny basin and advanced northeastwards against the eastern side of the Long Mynd towards Marshbrook and Little Stretton. The limit of penetration of Welsh ice in this area is not known with any certainty, but it probably extended as far as Little Stretton, where it came into contact with a lobe of the Irish Sea ice.

## IRISH SEA ICE

The drifts deposited over the northern slopes of the Long Mynd, in the Vale of Stretton and over the lowland area extending towards Shrewsbury, are very different in character from those found elsewhere. The till is typically brown or red-brown in colour, and contains distinctive erratics of Lake District and Scottish granites, fragments of marine shells picked up from the bed of the Irish Sea, and much Triassic material from the Shropshire Cheshire basih. The drifts are clearly the product of an Irish Sea ice sheet, a lobe of which moved southwards over the Shropshire-Cheshire basin, and invaded the Welsh Borderland hill areas on the west. Whitehead[1] considered that 'Irish

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1. Pocock, R.W., et. al., 1938, p. 182.

Sea' drifts are confined to the east of a line running from a point about 2½ ml. north-west of Shrewsbury towards Longden Common, and that to the west the drifts are of 'Welsh' origin, the product of an ice-stream moving down the Tanat, Vytnwy and Severn valleys.

The concentration of northern erratics between Lyth Hill and Baysten Hill, and the north-south orientation of Striae on Sharpstone Hill indicate that Irish Sea ice invaded this area from the north, and moved towards the Long Mynd and Vale of Stretton. On the Long Mynd, Irish Sea drifts extend up to about 1000 ft. and northern erratics are scattered over the hillsides above, up to 1300 ft. on Wilderley Hill. The dividing line between Welsh and Irish Sea drifts appears to run south from Longden Common, by Church Pulverbatch and onto Wilderley Hill. While no northern erratics are found west of this line, several from the west are found to the east of it. These include the Stiperstones Quartzites on Wilderley Hill, and the ryholite from Pontesford Hill at Underhill Hall. Thus, although the Irish Sea ice never extended west of this line, the Welsh ice evidently advanced further east, producing a maxing of the drifts of the two ice sheets.

The Irish Sea ice pushed southwards down the Vale of Stretton, depositing the characteristic red-brown drifts over the valley floor and adjacent hillslopes and leaving scattered erratics up to 1000 ft. on the hillsides in the north. It is not clear precisely how far the Irish Sea ice advanced down the Vale of Stretton. A borehole at Little Stretton showed 'solifluction gravels' overlying outwash gravels, to

a depth of 111 ft. but no trace of glacial till. This suggests that Little Stretton lay near the limit of ice advance. To the south and south-west of Little Stretton the drifts are of Welsh origin. However, patches of till to the south, at the Hough and Swiss Cottage, have yielded a few northern granites suggesting that this marked the limit of penetration of the ice.

## EXTENT OF THE ICE

In assessing the maximum extension of an ice-mass during a glacial episode, consideration must be given to the upper limits of the ice as well as to its lateral extent.

Both ice sheets and valley glaciers often deposit terminal moraines at their fronts, and the outermost moraine will give some evidence of the former lateral extent of the ice. Terminal moraines occur, most commonly, as an arcuate ridge or complex series of ridges extending across a valley, and are composed predominantly of till. They can only form effectively when the glacier is advancing or stationary in position and when the ice is active. Terminal moraines can also form during pauses in the general retreat of glaciers, or they may mark the limit of a minor re-advance. The presence of a moraine does not, therefore, necessarily indicate the maximum extent of an ice-mass.

At the terminus of some glaciers and ice-sheets a complex assemblage of kames and kettle-holes, or 'kame moraines', may be present. Such features are associated with deposition at the stagnant margin

of an ice-sheet during deglaciation, and not with active ice as are till moraines. Likewise, they may also mark a halt in the general retreat of an ice-sheet.

There are no true till moraines within the research area. Nowhere is there an arcuate till ridge extending across a valley. There are, however, several kame moraines, as for example across the Onny valley at Lyndham Heath, and in the Mule-Caebitra valley at Sarn.

Charlesworth associated the kamiform topography across the Onny valley with part of the end moraine of Welsh ice of Newer Drift age:

> "The moraine which floors the valley west of Craven Arms as far west as Clun, threads its way into the recess north of that locality, and flanks the hills near Kempton and Lydbury North and east of Bishop's Castle. It is spread over the bottom of the depression east of Lydham Station as far east as Eaton, while occasional mounds near Plowden suggest that the ice at this period attained the southern end of the Longmynd".[1]

However, this drift topography can in no way be regarded as an ice-limit, since extensive deposits of Welsh till extend continuously beyond the moraine into the Long Mynd - Shelve area. In all probability it represents an accumulation at or near the ice margin during a halt in the general deglaciation, Similarly, the kame moraines at Sarn, to the east of Montgomery, and on the northern slopes of the

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1. Charlesworth, J.K., 1929, p. 349.

Long Mynd do not indicate the maximum extension of the ice, but were formed during a period of ice retreat.

A possible terminal moraine of the Irish Sea ice sheet occurs in the Vale of Stretton. At the southern end of Brockhurst hill an arcuate ridge extends across the valley and is composed of dirty, poorly-sorted fluvio-glacial gravels. That there was an ice limit in this area is supported by the drift succession revealed by borehole records. To the north, at Church Stretton, up to 41 ft. of till was proved beneath the valley floor, while south of the Brockhurst moraine the thickness is reduced to some 15 ft., and at Little Stretton it is completely absent. With the decrease in the thickness of the till there is a progressive increase in the amount of fluvio-glacial sands and gravels, which were probably deposited while the ice-front stood a short distance to the north, at Brockhurst. The evidence suggests, therefore that the Brockhurst moraine marks a limit of the Irish Sea ice in this area. To the south of Little Stretton the drifts are very different in character and are referable to a Welsh ice mass moving into the area from the south.

Two distinct ice sheets entered the region from different directions and deposited till and fluvio-glacial deposits in all the valley floors and depressions. The two ice-sheets came into contact with each other on the northern slopes of the Long Mynd, and the line of contact can be readily distinguished by the contrasting nature of the glacial deposits on either side. The high ground between Cothercott Hill and Longden Manor appears to have been the meeting place between the two ice-sheets, though the actual line of contact fluctuated.

The Irish Sea ice did not extend into the Habberley valley to the west, though the Welsh ice appears at time to have advanced east and mingled its drifts with the ice from the Irish Sea basin.

Several lines of evidence may be used to determine the upper limits of the ice sheets. The most important include the limits reached by drift deposits and glacial erratics, the upper limits of glacial meltwater channels, and the recognition of landforms which may be considered as typical of unglaciated enclaves, for example tors.

Ithas long been recognised that the upper limit of continuous till cover does not necessarily mark the maximum extent of an ice mass. Isolated patches of drift and scattered erratics, referable to the same glacial episode, may extend beyond the limit of continuous drift. Numerous examples of this phenomenom are to be found in areas of comtemporary glaciation.

> "In almost all areas of present-day glaciation it is a matter of common occurrence to find bare rocky tracts, known to have been covered by ice not long ago, on which there is nothing but an occasional erratic to mark the event."[1]

The fact that in almost all areas of Pleistocene glaciation drifts

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1. Lamplugh, G.W., 1920, p. lxxv..

tend to be concentrated in the valley floors, and are only patchily distributed on the higher hillsides, is probably related to the original distribution of drifts within the ice-sheets. The ice sheets were predominantly cold, and in the present-day equivalents, such as those of Greenland, drift tends to be concentrated in the lower parts of the ice, while the upper parts are relatively clean.

> "Much of the debris load of glaciers and ice-sheets appears to be concentrated in bands in the basal part of the ice."[1]

> "The debris belts are essentially parallel to the base of the glacier. They are chiefly confined to the lower 50 or 70 ft., sometimes they prevail up to 100 ft. and rarely beyond."[2]

It is only when the ice ablates, or where the debris is shearing up at the margin, that the drifts become concentrated at the surface. Since the upper layers of the ice-sheet are relatively clean, it is logical that little drift is laid down on the higher hillslopes, and that the deposits are concentrated near the valley floors.

Luckman, working to the south of the research area, believed that scattered drift deposits and erratics at higher levels are remanee deposits of an earlier glaciation of much greater extent.[3]

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1. Boulton, G.S., 1967, p. 718.
2. Chamberlain, T.C., 1895, p. 203.
3. Luckman, B., 1966, p. 101.

This, however, seems unlikely, and he found no independent evidence of an interglacial episode. Undoubtedly, some deposits must have remained from an earlier glaciation, but these were probably re-worked during the last glaciation.

The distribution of glacial meltwater channels provides additional evidence on the upward extent of ice-sheets. This is particularly the case where the true marginal or subaerial nature of the channel is indicated. Where, however, the channels are submarginal or sub-glacial one cannot say that these mark the limit of the ice, which must have extended to higher altitudes. Even marginal channels do not necessarily mark the maximum encroachment of the ice.

> "It is generally accepted that marginal drainage channels will only operate below the snow-line, for above that level the cross-section of the land-ice contact will be concave upwards due to marginal snow accumulation and to a more rapid outflow of thicker ice some distance from the ice margin."[1]

Although meltwater channels cannot give a true indication of the maximum extent of an ice-mass, they are valuable indicators of the minimum level reached by the ice.

It may be argued that the higher meltwater channels are referable

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1. Hollingworth, S.E., 1952, p. 107.

to an earlier glacial expisode of greater extent than the last. Although it is likely that earlier glaciations will have left meltwater channels, it is very difficult to state that the higher channels are necessarily older. In this area no interglacial deposits were found, and the higher channels appear as fresh in form as the lowest. There is often an interconnecting system of channels from the highest hillslopes to the valley floor. If the higher channels are older, it seems likely that they have been re-occupied by meltwaters of the last glaciation. The meltwater channels are therefore considered to be of the same age.

In the extreme west of the research area, glacial till is patchily distributed along the high watershed of the Kerry Hills, between the drainage of the Mule and Teme and that of the Ithon. The till occurs up to 1550 ft. at Crugynnau, and attains its maximum thickness in the floors of cols in the watershed, as at Black Gate. Erratics within the drift show that the ice came from the west.

To the north-east the ice passed down the major valleys of the Kerry Hills into the Mule-Caebitra depression. The continuous drift cover does not extand much above 750-800 ft., but scattered deposits are found up to 1.200 ft. on the northern slopes of the Kerry Hills. Meltwater channels also occur well above the limit of continuous drift.

West of Sarn only isolated channels occur, as at Upper Wig and north of Cae-betin, up to 1,200 ft. East of Sarn, a complex system of channels extends up to over 1,400 ft. on the Kerry Hills. These are fresh in form and are little-modified by periglacial activity. They occur as a continuous series of marginal and submarginal channels and chutes, almost to the floor of the Caebitra valley. It is probable that they were utilized by meltwaters during the last glacial episode, and indicate that the ice reached up to at least 1,400 ft. in this area.

Welsh ice also entered the area from the west along the Severn valley. Here again, the continuous drift cover is confined to within about 100 ft. of the valley floor, but isolated deposits up to about 800 ft. occur on the Cefn-y-Coed hill mass to the south. There are few meltwater channels within this area, but the widespread development of streamlined, ice-moulded topography indicates that a substantial ice mass passed over the area. The ice-moulded landforms are not confined to the lowest ground, and even the highest peaks, up to 1,000 ft., show signs of modification. Linton believes that such forms must be produced beneath a thick and swiftly moving ice-mass, and that:

> "Such topography must surely require a cover of some hundreds of feet for its production."[1]

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1. Linton, D.L., 1957, p. 40.

Estimates vary as to the minimum thickness of ice beneath which such features are formed. Linton quotes a figure of 10 times the amplitude of the rock drumlins[1], while Fairchild states that the ice would be 250-500 m. above such features.[2] Since the amplitude of the rock-drumlins at Weston, at 1,000 ft., is about 50 ft. the ice surface was probably at 1,500 ft. over the Cefn-y-Coed hills; Fairchild's estimate gives a figure of 1,750 ft. These are undoubtedly estimates only, but the first figure agrees well with the height of the ice surface on the Kerry Hills deduced from the meltwater channels and glacial till.

The occurrence of drift drumlins within the Mule-Caebitra depression also strongly suggests that a considerable thickness of ice covered the area. Linton believes that a minimum ice thickness beneath which drumlins are formed is about 5 times the amplitude[3]; while Fairchild estimates that the ice surface would have been about 700 ft. above drumlins.[4] On these estimates, the ice surface would have been at about 950 ft. or 1,400 ft. within the depression. The latter figure again compares favourably with estimates from drift and meltwater channel distribution on the Kerry Hills.

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1. Linton, D.L.
2. Fairchild, H.L.
3. Op. cit.
4. Op. cit.

Glacial drift covers the floor of the Montgomery-Chirbury gap, and extends up to about 700 ft. against the steep western edge of the Corndon - Shelve hill mass. There is evidence, however, that the ice reached up to greater heights. There is an isolated deposit of drift in the col at Pant, above Old Church Stoke, at 800 ft. Till also occupies the floors of broad valleys on either side of Corndon Hill, up to about 1,000 ft., and it is likely that ice entered the Shelve area from the west along these two valleys.

Several ice-moulded ridges occur to the east of Chirbury, the largest of which is The Ridge at 800 ft. Using Linton's and Fairchild's estimates, the ice surface was probably at 1,300-1,500 ft. This, however, is only a crude estimate, and there is a lack of definite evidence that the ice reached much above about 1,100 ft. against Corndon Hill.

To the north-east, the Welsh ice moved along the Rea Brook - Camlad and Severn valleys towards the Shropshire Plain, being divided into two main streams by the Long Mountain. In the former valley the till is largely confined to the floor, and does not extend much above about 400 ft. Similarly, in the Severn valley, it does not reach much above 400-500 ft. Isolated patches of drift are found to much greater heights over the Long Mountain, and meltwater channels occur well above the limit of continuous drift.

On the Long Mountain, isolated patches of drift are found at all

elevations, often in cols, depressions and valley floors, right up to the watershed at 1,325 ft. East of the watershed, at Beacon Ring, glacial deposits are exposed in numerous small sections at the head of the Mulsop valley and on neighbouring hillslopes. Similar deposits are found in the floors of valleys draining eastwards from Welsh Harp, and are common on hillsides and in valley floors around Rowley, Heldre Hill and the Wallop valley. Meltwater channels are also found right up to the watershed, to a maximum of 1,275 ft. Complex channel systems occupy the valley heads to the east of the watershed; and several marginal and submarginal channels occur on the sides of the larger valleys, as at Upper Monksfields (1,000 ft.) and in the Mulsop valley (900 ft.). A prominent channel cuts directly across the interfluve between the Mulsop valley and Outrack Dingle at 1,050 ft. These channels are all fresh in form, and appear to have been utilized by meltwaters during the last glacial episode. The distribution of meltwater channels and associated patches of till indicate, therefore, that the ice surface probably covered the Long Mountain during the last glaciation, reaching up to at least 1,325 ft.

Welsh ice moved eastwards from the Church Stoke area down the Camlad - Onny valley towards Lydham and Plowden, where it was joined by an ice-stream passing north-eastwards towards the Long Mynd from the Clun valley.

Drift deposits are again largely confined to the valley floors, and do not extend much above 600 ft. in the Camlad valley and 700 ft. in the Onny valley. On the southern side of the Camlad valley, however, isolated drift deposits and meltwater channels extend up to high levels on Aston Hill. The drift is largely confined to valley floors, for example above Pentre; several erratics of fine-grained igneous rocks were found on Aston Hill at 1,200 ft., and at patch of till above Whitcot Stile is at 1,150 ft. Meltwater channels are concentrated on the hillslopes above Pentre, where a series of submarginal channels is at about 800 ft., and the highest intake of the subglacial chutes is at 900 ft. It seems likely, therefore, that the ice extended up to at least 1,200 ft. against Aston Hill, which may in fact have been completely covered.

East of Bishops Castle the hills are devoid of drift, and no erratics were found at high levels. There are meltwater channels up to 800 ft. on Oakley Mynd. To the south of the river Clun, Plynlimon erratics occur up to 1,400 ft. on Black Hill, and in view of the height of the till in the Long Mynd - Shelve area to the north, it seems probable that hills such as Oakley Mynd were covered by ice, though little trace remains.

On the northern side of the Onny valley patches of drift are found up to a height of over 1000 ft. on Heath Mynd and Norbury Hill, above which are scattered erratics up to the summit of the latter, at 1325 ft.

The Welsh ice pushed northwards up the East Onny valley and up the broad valley between Cefn Gunthley and Corndon Hill to invade the Shelve-Long Mynd hill area. The upper hillslopes are largely drift-free, but scattered erratics testify to a former ice cover. These occur, for example, to the south of The Bog up to about 1,200 ft., and to about 1,100 ft. south of Shelve. Meltwater channels are also to be found at these high levels. East of Shelve a channel cuts across a watershed at 1,125 ft., and the intake of the Ritton Castle channel is at 1,025 ft., Marginal and submarginal channels north of Corndon Hill are between 1,000 and 1,100 ft., and the intake of the prominent Hope valley channel is at 1,000 ft. on Black Marsh. In the Shelve area, therefore, there is evidence that the ice stood up to at least 1,100-1,200 ft.

Though the Stiperstones Ridge is largely drift-free, meltwater channels occur to even greater elevations than those in the Shelve area. They are concentrated chiefly on the northern end of the Ridge, and reach a maximum elevation above Snailbeach and the Crowsnest Dingle. Above the former, the intake of the Eastridge channel is at 1,100 ft. at Lordshill. To the south, a series of subglacial chutes plunge into the eastern side of Crowsnest Dingle and their intakes are at 1375 ft. at Blackmoorgate. It is reasonable to conclude that the ice oovered the northern end of the Stiperstones up to about 1,400 ft.

The higher, central portion of the Stiperstones is free of both channels and glacial drift, being dominated by tors and periglacial deposits. However, a group of small channels plunges eastwards from

a low col in the ridge at The Knolls, and the intakes of these are at about 1,400 ft.

To the east of the Stiperstones Welsh ice penetrated northwards up the East Onny valley, southwards up the Habberley Brook valley, and the upper ice layers moved eastwards across the Stiperstones. Both of these large valleys are floored with drift, which forms a continuous cover up to 1,100 or 1,200 ft. in the valley heads, as at Gatten Lodge. Though the hillsides adjacent to the East Onny valley are covered with substantial thicknesses of head, isolated patches of drift and scattered erratics occur well above the limit of continuous drift. Several erratics of rhyolite and quartz conglomerate from the west and north-west occur at Medlicott, at 1,050 ft. Similarly the eastern side of the Habberley valley has occasional patches of drift in cols and depressions.

The northeastern part of the area is dominated by the effects of a branch of the Irish Sea ice sheet, which moved southwards from the Shropshire-Cheshire Plain up onto the northern slopes of the Long Mynd and into the Vale of Stretton.

On the northern slopes of the Long Mynd, the upper level of continuous drift is at about 1,100 ft. in the west, and decreases to about 700 ft. in the east. However, there is substantial evidence, from scattered drift deposits, erratics and meltwater channels, that the ice reached to greater elevations.

Numerous erratics are scattered over the upper slopes of Cothercott Hill and the adjacent Wilderley Hill, up to heights of 1,350 ft.

An erratic of Eskdale granite was found at 1,300 ft. at the Thresholds. Several large meltwater channels also have their intakes on these upper slopes; for example, at the Thresholds, at 1,325 ft., and on Cothercott Hill at 1,250 ft. One channel, Golden Valley, to the southeast of the Thresholds is located in the floor of a col at 1,250 ft. in the Long Mynd watershed. Farther east, several channels have intakes at similar elevations; for example at High Park House and near Dry Hill.

Within the Vale of Stretton, the drift is largely confined to the valley floor, very little occurring on the hillslopes above 800 ft. Isolated patches of till are to be found at Bullocks Moor Cottage and Plush Hill, at about 1,050 ft. on the Long Mynd, and scattered northern erratics are common in the adjacent area. On the opposite side of the valley granite erratics were found at over 1,000 ft., on the northern end of Caer Caradoc and on The Wilderness. A complex system of meltwater channels has also been eroded into the hillslopes above All Stretton, including a prominent sub-marginal channel at Worsely, at 1,000 ft.

From the form and position of the meltwater channels within the Vale of Stretton it is possible to infer the slope and extent of the ice surface. The highest channels are in the north, above All Stretton and on Little Caradoc, at about 1,000 ft. To the south the submarginal and marginal channels progressively decline in elevation, to about 800 ft. above Church Stretton and 700 ft. at Brockhurst Hill. These heights compare favourably with the upper limits of drift and erratics on the opposite side of the valley; about 800 ft. opposite Church

Stretton and 650 ft. to the south.

It would therefore appear that the surface of the branch of the Irish Sea ice sheet was up to at least 1,050 ft. in the north of the Vale of Stretton and at about 700 ft. at Brockhurst. A possible terminal moraine of this ice has been deposited against the southern end of Brockhurst Hill. The steep-sided valleys, or 'batches', which drain the eastern slopes of the Long Mynd appear to have been hardly penetrated by any glacier ice. Drift deposits are found only at their mouths, and they are largely devoid of meltwater channels. Considerable deposits of head and scree mantle their slopes. The fact that the valleys are orientated transverse to the ice movement largely explains why they were never invaded by ice.

South of Little Stretton the glacial drifts were deposited by a 'Welsh' ice sheet moving north-eastwards up the Horderley-Marshbrook depression. Again, the hillslopes above about 800 ft. are drift-free, and there is little evidence of glacial action above this height. The area is close to the limit reached by the Welsh ice, and one would expect a lower ice surface in this position.

Unglaciated Enclaves

Another way to approach the problem of determining the limits of ice-sheets is by the recognition of 'unglaciated enclaves'. It is thought that unglaciated areas should possess specific characteristics of surface and form which should permit them to be recognized as such. At the end of the last century workers in Scandinavia

proposed that several plants probably survived throughout the Ice Age in unglaciated areas, and they developed the 'nunatak hypothesis'. Since then a continuing controversy has developed, although more data has been accumulated and other criteria cited. In particular, a considerable controversy has surrounded the significance of tors as indicators of unglaciated enclaves.

In Britain, Linton is the chief advocate of tors being characteristic of unglaciated areas:

> " ... true tors are one of the group of associated features... by which we may hope to recognize surfaces that have survived from interglacial, or, more probably, pre-glacial times."[1]

Linton formulated his theory in the Dartmoor area and implied that tors originated in two phases. Firstly, there was a period of deep weathering of the granite, most probably under the influence of wet, tropical climate in Tertiary times. Subsequently, the tors were exhumed by the removal of the weathered debris.[2] Both phases probably predated the glacial period. Tors in many other parts of Britain and Europe were explained by Linton in a similar manner.

Other workers, however, while agreeing that tors are indicators of climatic change, have cast doubt both on Linton's theory regarding their origin and on their role as indicators of unglaciated enclaves.

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1. Linton, D.L., 1949, p. 32.
2. Linton, D.L., 1955, and 1956.

Palmer and Radley[1], for example, maintain that the gritstone tors of the Pennines were formed exclusively in a periglacial environment by gradual evolution. Palmer and Nielson[2], think that the granite tors of Dartmoor evolved in a similar manner. The tors are not 'core stones', there is no Tertiary regolith, and they have arisen without the action of deep weathering. The action of frost on exposed rock surfaces is concentrated in joints and other lines of weakness, and the break-up of such a bedrock by frost produces blockfields which gradually move downslope by solifluxion.

> "The relationship of tor, shattered blocks, blockmeer and tor, free face, landslides and solifluxion emphasises that tors are a by-product of the much larger scheme of slope evolution under periglacial conditions, and if any minor Tertiary landforms suchas a corestone were to be found on the Pennines it would be totally unrecognizable."[3]

Tors could thus be formed in extra-glacial areas by periglacial activity during the Pleistocene. Tors which exist in areas believed to have been glaciated could also have been created under periglacial conditions during the retreat from the ice maximum.

More recently Dahl[4] has studied tors in northern Norway and contends that both frost weathering and chemical weathering may combine to produce

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1. Palmer, J and Radley, J., 1961.
2. Palmer, J. and Nielson, 1962, p. 315-319.
3. Palmer, J. and Radley, J., 1961, p. 46.
4. Dahl, R., 1966. p. 78-84.

tor-like forms, and that these are present in areas covered by ice during the last glaciation. He concludes that tors should be used with the greatest caution in evaluating the extent of Quaternary glaciations, and they cannot be used as evidence of unglaciated areas.

In south-west Shropshire a prominent group of tors rise abruptly above the convex upper slopes of the Stiperstones Ridge (Pl. 38). They are confined to the higher parts of the ridge, above 1,400 ft., and are composed of a tough Arenig quartzite which dips from 40-60° towards the north-west.

> "There is not, perhaps, a more singular feature in the physical geography of England than the Stiperstones. These rocks are made up of a number of broken and serrated ledges, jutting out to form the summit of the hills ... They stand out on the crest of the ridge at short intervals, like rugged cyclopean ruins, some of the principal of which are about 50-60 ft. high, and about 120-130 feet in width....The main range of the serrated points is from NNE-SSW, and therefore conformable to the strike of the adjoining district of Shelve; but when examined in detail, this ridge is found to be intersected by a number of transverse faults, dividing it into several masses..."[1]

The tors are formed of well-jointed quartzites, and the joints and bedding planes have been picked out by physical, and perhaps some

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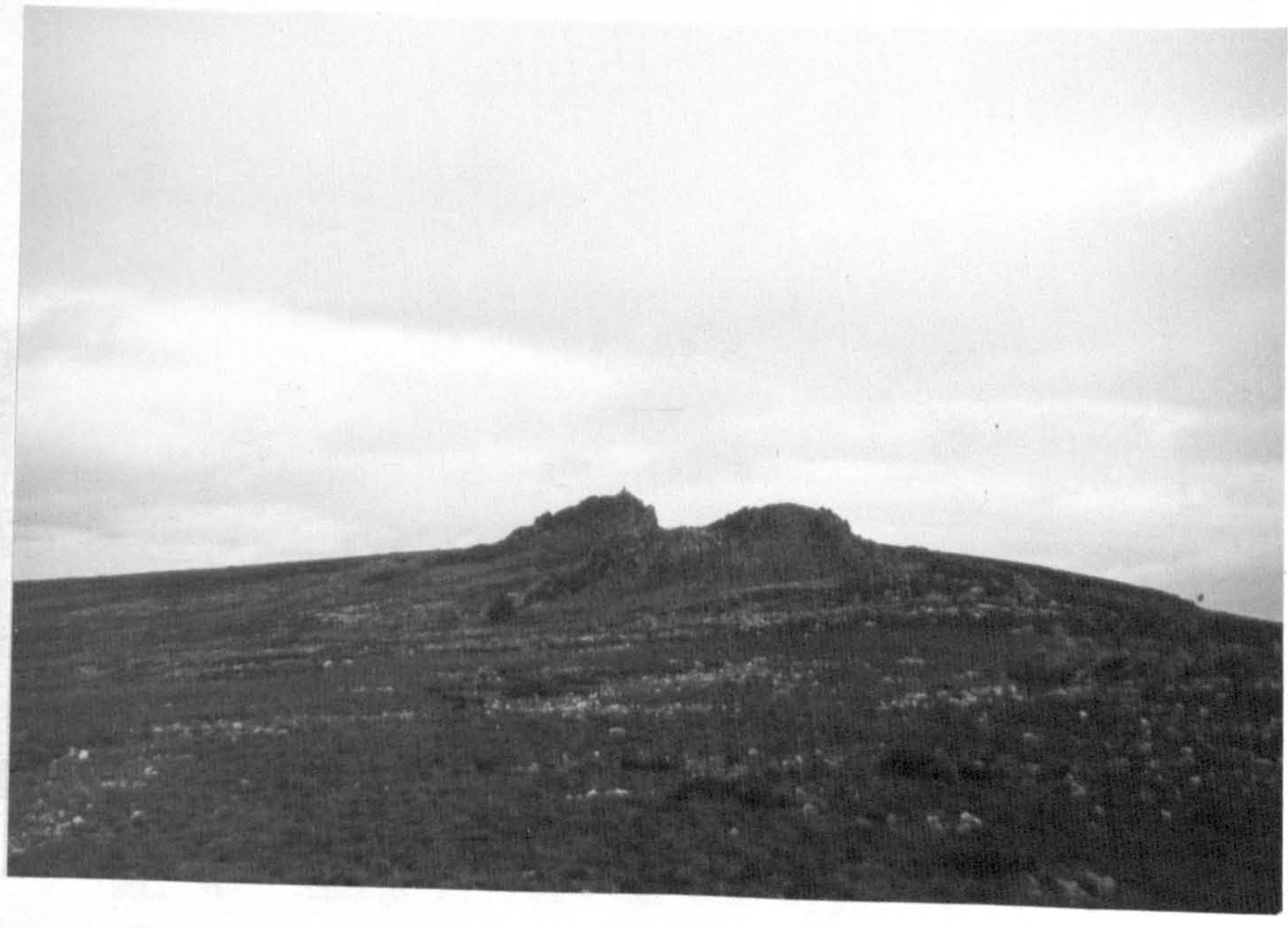
1. Murchison, R.I., 1839, p. 283.

chemical, weathering. Blocks of angular quartzite have evidently fallen from the tors onto the surrounding hillslopes. The slopes of the surrounding moorland are strewn with angular blocks of quartzite, often forming blockmeer (Pl.39 ). The blocks are generally from 6-8 in. across, and are apparently derived from the destruction of the tors. The latter are found in varying stages of destruction and some, such as Shepherd's Rock and Scattered Rock (Pl.40 ), have almost disappeared.

Stone patterns, in the form of polygons and stripes, can be picked out in the blockmeer ( Pl.41 ). Though many large loose blocks have rolled down on both sides of the ridge, they are in much greater abundance on the eastern face. The processes of destruction and mass-movement do not appear to be active at the present day. Some of the lower, more destroyed tors, such as Shepherd's Rock, are partly covered by vegetation, and much of the blockmeer, stone polygons and stone stripes are also vegetated and stabilized.

It seems unlikely that the Stiperstones tors originated in the manner envisaged by Linton. There is no trace of any deep weathering regolith, and the tors cannot be regarded as core-stones; there are no smooth, rounded boulders as found by Linton on Dartmoor. It is more probable that they were formed under the influence of periglacial activity during a cold climate. Frost action, aided by the presence of numerous joints and bedding planes in the quartzite, would act on the rock, preparing it for mass-movement and solifluxion. Softer beds outcrop on either side of the quartzite, and the quartzites themselves

Pl. 38 Devil's Chair, Stiperstones



Pl. 39 Mainstone Rock, Stiperstones

Pl. 40 Shepherd's Rock, Stiperstones

Pl. 41 Stone stripes on the upper slopes of the Stiperstones Ridge

are probably of differing hardness. The softer bands would be worn away first and the harder quartzites remain as long projecting ridges, or tors, along the crest of the hill range. Large quantities of detritus and rubble moved down the hillslope by solifluxion. Stone patterns remain on the upper slopes, but the lower slopes, particularly around Gittinshay Wood and above Gatten Plantation, are mantled with a considerable thickness of solifluxion material. It would appear, from the height of the tors, that the surface of the Stiperstones Ridge has been lowered by up to 60 ft. by periglacial activity.

The process of formation appears to be no longer active, and the tors must have originated in a climate more severe than the present, probably during the glacial epoch. Andersson[1] maintains that solifluxion is the chief agent of destruction in a subglacial climate, with heavy deposits of winter snow melting in summer.

> "... the subglacial climate must be looked on as an optimum of destructive action."[2]

The relationship between the Stiperstone tors and the upper limits of glaciation remains somewhat problematical. It is possible that they were formed in an unglaciated enclave, or nunatak, during the last glaciation. Evidence from the upper limits of drift, and, more especially, of meltwater channels, suggests that the ice did not reach much above 1,400 ft., and it is at these higher elevations that the tors are found. However, meltwater channels do not mark the maximum upper limit of the

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1. Andersson, 1906, p. 110.
2. Op. cit. p. 112.

ice, only a minimum level, and it is possible that the ice reached higher elevations. Dahl[1] has shown that tors can occur in areas that were glaciated during the Wurm and that they cannot be used as evidence of unglaciated enclaves. It is possible that the Stiperstones were completely covered by glacier ice, and that during deglaciation the higher areas would be uncovered first, and would exist as nunataks surrounded by ice. The tors could then evolve in this extra-glacial environment during glaciation. Evidence elsewhere also suggests that considerable periglacial activity continued during the Late Glacial period. In the floor of the Vale of Stretton there are considerable thicknesses of periglacial gravels and clays, which Rowlands[2] has shown to be of Zones 1 and 111 age. It is not unlikely that tor formation continued on the Stiperstones during the Late Glacial.

Although there is no firm evidence to indicate that the Stiperstones Ridge was completely ice-covered, the possibility cannot be entirely ruled out, and the tors themselves are not positive criteria of an unglaciated enclave.

## CONCLUSION

During the Last Glaciation south-west Shropshire, in common with other areas of the Welsh Borderland, was invaded contemporaneously by two separate ice sheets. One was nourished in the mountains to the

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1. Dahl, R., 1966, p. 84.
2. Rowlands, P.H., 1966, p. 69-80.

west, in the Plynlimon region of central Wales, while the other came from the Irish Sea basin, passing southwards across the Cheshire-Shropshire Plain.

The glacial till of the western ice is typically brown or blue-grey and includes erratics derived from the rocks of the Central Welsh Syncline. Particularly significant among these are the ' Plynlimon ' grits of Rhayader and Llandovery which have been traced in a west-east line from Beguildy to Clunbury and on the Kerry Hills, at all elevations up to 1400 ft. This distribution clearly demonstrates a general eastward movement of the upper layers of the ice sheet from central Wales. A similar ice movement has been elucidated by Dwerry-house and Miller[1] and by Luckman[2] in the country to the south, almost as far as Wye valley. (fig. 6.1).

Welsh ice entered south-west Shropshire along two main routes (fig. 6.1). A considerable volume crossed the watershed between the drainage of the Teme and Mule and that of the Ithon. At lower levels the ice moved north-eastwards down several large valleys to converge in that of the Mule near Kerry; from here it passed along the Mule-Caebitra depression towards Church Stoke. At higher levels the ice moved generally eastwards along the northern edge of the Kerry Hills. The distribution of scattered till deposits and meltwater channels testifies that the ice reached up to about 1400 ft. against the Kerry Hills.

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1. Dwerryhouse, A.R., & A.A. Miller, 1930.
2. Luckman, B.H., 1966.

A second important stream of western ice entered the region along the broad Severn valley, moving northeastwards towards Forden. The extensive ice-moulded topography and the scatter of till deposits indicates that the Cefn-y-Coed hill was completely ice-covered at this time.

There is evidence that Welsh ice entered the region along a third route, from the Clun Forest area. West of the Long Mynd, on the lower slopes of Norbury Hill and scattered up the East Onny valley are several Wenlock Limestone erratics. A northward movement of ice up the East Onny valley is also indicated by the distribution of erratics of Pentamerus sandstone. The limestones were presumably derived from outcrops to the south, and were introduced through the two broad gaps on either side of Oakley Mynd. It is possible that pressure of ice from the Teme, Lugg and Arrow valleys to the south caused ice streams from the Teme and Clun to move north and northeast towards Craven Arms and the East Onny valley, as suggested by the Soil Survey[1]. This ice was confluent with ice advancing down the Camlad valley and the combined streams apparently moved around the southern end of the Long Mynd towards Marshbrook and up the East Onny valley towards Bridges.

In the western part of the region the movement of the Welsh ice was greatly influenced by the topography, and the principal ice streams passed down the several broad valley lines. That moving along the Severn valley bifurcated against the southern end of the Long Mountain;

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1. Mackney, D., & C.P. Burnham, 1967, p. 15, fig. 7.

one stream continued northeast towards the Breiddens, and the other headed along the Camlad-Rea Brook valley towards Minsterly. It is clear that the Long Mountain itself was occupied by ice, and the occurrence of till deposits and meltwater channels right up to the watershed at 1,325 ft. suggest that it was probably completely covered.

Ice moving down the Mule-Caebitra depression bifurcated in a similar manner against the southern flank of the Corndon hill. One branch was deflected northwards towards Marton, while the other passed east and south-east down the Camlad-Onny valley towards Plowden.

Ice movements within the Corndon Hill-Stiperstones-Long Mynd hill area are rather more complicated. Ice advanced into the region from the south, usually along the larger valleys. There also appears to have been a movement of the upper layers of the ice across the strike of the hills, from west to east. Erratics of Stiperstones quartzites are common in the East Onny valley, on the eastern side of the Habberley Brook valley, and on Wilderley Hill. Finally, that ice passed southwards up the Habberley Brook valley is shown by the distribution of erratics of Pentamerus sandstone.

Glacial till is generally confined to the valley floors in this region, but the distribution of erratics and meltwater channels demonstrates that the ice reached up to a higher elevation. Channels are found up to 1,400 ft. on the Stiperstones, and indicate a minimum level reached by the ice. The Stiperstones tors are not conclusive evidence of nunatak conditions during the Last Glaciation since they could have been formed in a periglacial climate subsequent to ice retreat.

On the other hand, there is no firm evidence that the ice reached much above 1,400 ft., either on the Stiperstones or on Corndon Hill.

The north-eastern part of the district is dominated by the effects of an ice sheet which advanced southwards from the Irish Sea basin. Part of the ice sheet passed southwards by Shrewsbury, extended onto the northern slopes of the Long Mynd, and sent a branch up the Vale of Stretton. The glacial tills are readily distinguishable from those of the western ice; they are typically brown or red-brown in colour, contain numerous erratics of Triassic sandstones, granites and other igneous rocks from the Lake District and Scotland, together with occasional fragments of marine shells.

From the distribution of scattered till deposits, erratics and meltwater channels it is evident that the ice reached up to at least 1,300 ft. on the Long Mynd. In the Vale of Stretton it seems that the level of the ice declined towards the south. In the north the ice was at about 1,000 ft., but it was probably only at about 800 ft. at Church Strettoh. The Brockhurst moraine probably marks the limit of the ice advance in this area.

The higher parts of the Long Mynd, up to 1,650 ft., possibly remained as a nunatak. There is no trace of ice action at these high levels and the surface is mantled with thick deposits of head. It is also likely that the upper parts of the large eastward-draining 'batches' remained ice-free, largely because they were orientated transverse to the general direction of ice-movement.

In the northern part of the region the distribution of distinctive erratics of Breidden Hill dolerite and Carboniferous Limestone

indicate a movement of Welsh ice from north-west to south-east, across the general northeasterly flow over the rest of the region. This transverse movement probably reflects the interplay of three separate ice streams.

To the south of Oswestry ice debouching from the Severn valley came into contact with that pressing eastwards along the Tanat and southeastwards from the Berwyn hills[1]. The pressure of the latter would tend to deflect the Severn ice to a more easterly flow. In pushing out onto the Shropshire Plain the combined Welsh ice streams would also come into contact with the southward moving Irish Sea ice. The outward pressure of the Welsh ice was evidently such that it prevented the latter from invading the area south of Oswestry. Here, and over the ground to the west of Shrewsbury, the tills are entirely of Welsh origin. The combined Welsh and Irish Sea ice streams would tend to move in a common south-east direction along their zone of contact, as a resultant of their opposing forces.

The zone of contact between the two ice sheets can be determined by the contrasting nature of their respective deposits. North of Oswestry the elevated outcrop of the Cefn-y-Fedw sandstone marks the western limit of the Irish Sea ice sheet.[2] No sign of the ice has been found west of a line runnihg through Selattyn Hill to Mynydd-myfeyr, Sweeney Mountain and Llanymenech Hill. Galcial gravels containing shell fragments and erratics of north-country rocks are found up to nearly 1,200 ft. along this escarpment.[3]

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1. For full details of these ice movements see Wedd, C.B., et. al., 1929; and D.C. Davis, 1876.
2. Wedd, C.B. et. al.
3. Nicholson, A.C., 1892.

South of Oswestry, however, there is no trace of the Irish Sea drifts:

> " ... the great efflux of ice from the Tanat, Vyrnwy and upper Severn valleys over ground apparently never reached by Northern ice, accounts for the exclusively Welsh type of the drifts there found."[1]

To the west of Shrewsbury the line of contact between the two ice sheets seems to have run by Longden and Longden Manor to Cothercott Hill and Wilderley Hill. No trace of the Irish Sea ice has been found to the west of the line of hills on the eastern margin of the Habberley Brook valley.

The actual zone of contact must, nevertheless, have been somewhat diffuse and there was inevitably a degree of intermingling of the two ice sheets. This would result in mixed drift deposits of northern and Welsh material in a broad belt at the junction of the two ice masses. The margin of one would undoubtedly at certain times override that of the other, depending on their relative pressures and thicknesses during the various stages of glaciation. The stratigraphical relationships of the drifts along this zone of contact would therefore not necessarily indicate a time interval between the advance of one ice stream and that of the other. The intermingling of the drifts is evident in the Oswestry area to the north. In the present area it is to be seen on the northern slopes of the Long Mynd, where Welsh erratics were found on Wilderley Hill and at Underhill Hall. While the Irish Sea ice seems to have been prevented from extending further west, the Welsh ice at times evidently advanced further east.

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1. Wedd, C.B., et. al., 1929, p. 162.

In the Welsh Borderland to the north there is a measure of agreement that The Welsh ice stream appears to have made its influence felt at a period when the northern ice had already, to some extent, established itself in the area. There is evidence that the first ice to reach the Oswestry district was from the north, but it was probably not long before the Welsh ice descended from the hills to deflect it. Since there is no trace of the no rthern drift west and south of the Carboniferous escarpment the Geological Survey maintains that the Welsh ice must already have occupied the ground, and that the first contact between the two was along the line of the escarpment.

A similar state of affairs is evident in south-west Shropshire. Since no trace of the Irish Sea ice is found west of the hills bordering the Habberley valley, or in the area south of Oswestry, the Welsh ice must have already been in existence when the Irish Sea ice advanced into the area.

# Chapter 7

## DEGLACIATION

"In all areas glaciated during the Pleistocene period, deglaciation features are often so abundant that they dominate the scenery. Studies of deglaciation are therefore of fundamental importance for our understanding of the genesis of a landscape in glaciated areas."

(Hoppe, G., 1967, p. 204)

## INTRODUCTION

In assessing the character of deglaciation in any area, consideration must be given both to the forms of fluvio-glacial deposition as well as to those of erosion. The latter have been discussed in some detail in Chapter 5, and attention will here be mainly concentrated on the fluvio-glacial deposits. Both the form of the deposits and their composition are significant in determining their origin. However, the lack of good exposures necessitates that attention be focused on the former.

> "In all questions relating to the origin of the drifts, the topographic forms which the deposits assume are of at least as great importance as their lithological composition."[1]

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1. Lapworth, G.W., 1903, p. 353.

As with meltwater channels, divergent views exist as to whether fluvio-glacial deposits are associated with glacier-dammed lakes, or with downwasting ice sheets and stagnant ice margins. Until relatively recently, the interpretation of fluvio-glacial deposits in Britain has almost always been influenced by the concept of normal horizontal retreat of the ice front rather than by the, now generally accepted, idea of glacier downwastage.

Since 1898, when Lapworth[1] first proposed that the upper Severn was added to the lower Severn as a result of the impounding of a glacial lake in the Shropshire basin, and that it drained across the pre-glacial divide at Iron Bridge eroding the well-known gorge, the idea that a large lake, named Lake Lapworth, occupied the Shropshire-Cheshire plain has been widely held. Harmer[2] later put forward the same view, basing his hypothesis on the Iron Bridge overflow channel and the existence of supposed lake deposits on the Shropshire plain. The latter include terraces of sand and gravel and laminated brick-earths. The numerous kames are explained by 'lacustrine denudation':

> "The occurrence of isolated mounds, often capped by gravel, in the area of the suggested lakes is a feature common to them all...While in some places inequalities would be levelled by the accumulation of sediment, in others, as the overflow channels were deepened and the lakes became shallower, what I venture to call 'lacustrine

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1. Lapworth, G.W., 1898, p. 425.
2. Harmer, F.W., 1907, p. 477-481.

denudation' might, I think, have taken place."[1]

One of the most detailed studies of lake Lapworth and associated deposits near the Iron Bridge gorge has been carried out by Wills.[2] (fig. 7.1). The advance of the Irish Sea glacier impounded a series of lakes which were only of a transitory nature. It was on the retreat of the ice sheet that the largest and most important lakes were formed and diversions of drainage resulted. The Iron Bridge overflow originated from a small ice-dammed lake in Coalbrookdale, Lake Buildwas. Further withdrawal of the ice led to the formation of lake Lapworth, which came into being when a lake around Newport was united with Lake Buildwas. Wills recognised several phases in the retreat of the Irish Sea ice. The ice stands are related to supposed ice-marginal deposits, especially kames and beaded eskers. In the Wolverhampton-Newport esker train 19 successive still-stands have been interpreted. It is estimated that Lake Lapworth stood at a general level of about 300 ft., based on the heights of supposed shoreline features, the height of kamiform deposits at the entrance to the Iron Bridge gorge, and the projected thalweg of the Main Terrace of the Severn, which is believed to have been formed at this time.[3]

Following Wills's work, the idea that a large ice-dammed lake, or series of lakes, occupied the Shropshire plain during the retreat of the ice of the Last Glaciation has remained very strong in the

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1. Op. cit., p.481.
2. Wills, L.J., 1924, p. 274-314.
3. See p. for a fuller discussion of the significance and dating of this terrace.

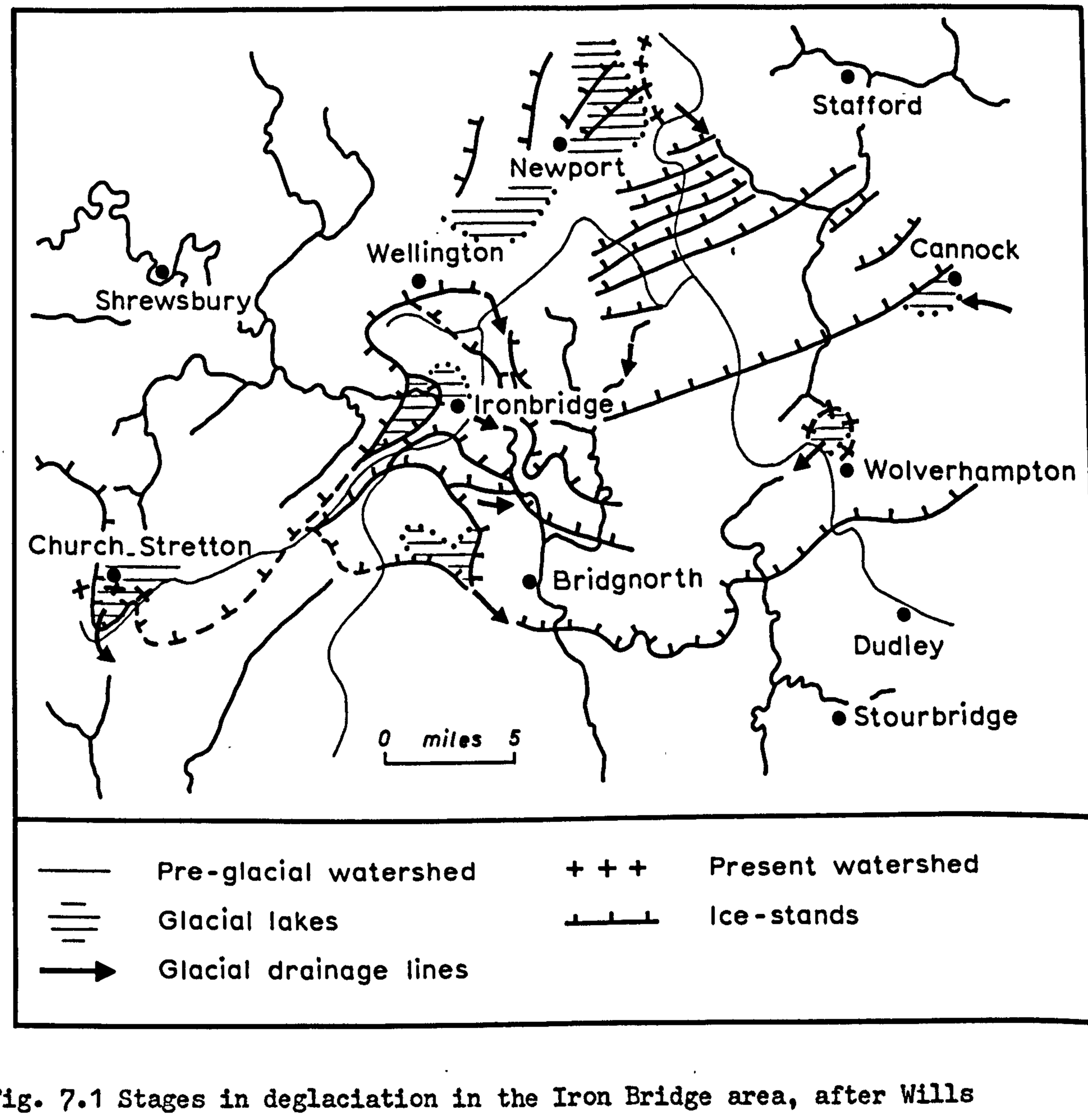


Fig. 7.1 Stages in deglaciation in the Iron Bridge area, after Wills

literature. Recently, Poole and Whiteman[1] extended the original limits of Lake Lapworth as defined by Wills. They considered that it covered the whole of the southern and eastern part of the plain and extended north into the Wigan district. They also recognised a higher, 330 ft., Lake Newport in the Wigan area, along the western margin of the Pennines, within the Nantwich district, and in the southern part of the basin generally:

> "Between Gnosall and Welshpool, around the southern margin of the basin, the level is only poorly developed, but near Welshpool it forms a good feature first described by Dwerryhouse and Miller.[2]

The 'good feature' refers to the level of the gravels within the Marton valley. Dwerryhouse and Miller[3] held that the Marrington Dingle 'overflow channel' had an outlet for its waters by way of the Marton valley and so to the Shropshire Plain; the western end of the valley was at this time occupied by ice. Since the level of the gravels within the valley is about 335 ft., the lake on the plain of Shropshire, Lake Lapworth, must have been below this. It has already been established that Marrington Dingle was eroded subglacially, and not as a lake overflow, and it seems doubtful that the meltwaters flowed into a lake in the Marton valley.

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1. Poole, E.G., & A.J. Whiteman, 1961, p. 113-118.
2. Op. cit., p. 116.
3. Dwerryhouse, A.R., & A.G. Miller, 1930, p. 102-103.

Wills also indicated a glacial lake impounded in the Stretton valley, overflowing through the Marshbrook gorge.[1] This lake was apparently contemporaneous with Lake Buildwas. It has been shown that the Marshbrook gorge was also formed subglacially, and that the meltwater channels within the Vale of Stretton are either submarginal or subglacial. It seems unlikely that a glacial lake occupied the valley.

A study of the character of deglaciation in the present research area will help throw considerable light on the validity of the former existance of Lake Lapworth and Poole and Whiteman's higher Lake Newport. The area borders on the Shropshire Plain, and if it can be demonstrated that the mode of ice retreat was by gradual downwastage and stagnation of the ice margins, and that extensive glacial lakes did not exist, doubt is cast on Lake Lapworth.

## RETREAT FEATURES OF THE IRISH SEA ICE SHEET

Within the Vale of Stretton there are many landforms which may be attributed to fluvio-glacial erosion and deposition. There is a complex system of meltwater channels concentrated on the lower slopes of the Long Mynd, and on the valley floor there are numerous kames and sands and gravel deposits (fig. 7.2).

The distribution, morphology and origin of the meltwater channels was discussed in some detail in Chapter 5. It was concluded that they were formed at or near the margin of a downwasting and stagnating ice

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1. Wills, L.J., 1937, p. 92, Fig. 6.

mass, and that the majority were cut submarginally or subglacially. At an early phase of deglaciation there was a concentration of meltwaters along the ice margin and the cutting of a system of submarginal channels along the edge of the Long Mynd. Many are best explained in terms of the superimposition of englacial streams from the ice onto the underlying topography consequent upon ice thinning. Most of the channels slope down towards the south, reflecting the general slope of the ice surface. At a later stage, the submarginal drainage was replaced by meltwaters plunging directly into the ice, often eroding subglacial chutes on the sides of the larger submarginal channels. In some cases the chutes plunge down valley sides from the intakes of earlier-formed submarginal channels. Finally, it would appear that the glacial drainage was appropriated along the main valley lines.

The fluvio-glacial material on the floor of the Vale of Stretton was deposited at a relatively late stage of deglaciation. Numerous kames, several rising up to 30 ft. above the flood plain, dominate the area between All Stretton and Church Stretton. In the large kame adjacent to Spring Bank Farm there are several shallow kettle holes.

Kames may form in a variety of ways, and their origin is often somewhat problematical. Flint considered that the presence of kettle holes is proof of stagnant ice, and they are usually found in the terminal zone of glaciers.[1]

> "A kettle represents either separation of a mass of ice or burial of ice beneath drift. In either case the ice must have

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1. Flint, R.F., 1957, p. 152.

> been thin...If thin and detatched, the ice must have been stagnant. If thin and buried, it must likewise have been stagnant...Hence we may conclude that kettles record ice masses that were not only thin but stagnant."[1]

Flint further states that shallow kettle holes are formed by the melting of relatively thin, buried ice masses.[2]

The occurrence of shallow holes within the kames in the Vale of Stretton suggests, therefore, that the latter were deposited in relatively thin and stagnant ice, probably at the glacier margin. The lack of bedding and the unsorted nature of the deposits, together with the lack of recognizable ice-contact slopes, argues against the kames being deposited as small deltas or fans at the ice margin. It is unlikely that they were deposited in a glacial lake. The kames probably originated as sediment deposited in crevasses or other openings in or on the surface of stagnating ice, which later melted away.

> "In a broad sense kettles are the counterparts of kames ... both features are the product of the ablation of thin glacier ice."[3]

The kame complex between All Stretton and Church Stretton was deposited at the stagnant margin of the Irish Sea ice sheet, at a relatively late stage of deglaciation. The smaller group of kames to the north, at Botvyle, which also have kettle holes in them, probably

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1. Op. cit., 1942, p. 126.
2. Op. cit., 1957; p. 151.
3. Ibid.

marks another ice margin, when the ice had retreated slightly upvalley.

It has been demonstrated that the Marshbrook gorge at the head of the Vale of Stretton is not an overflow channel of a glacially-dammed lake, but was eroded subglacially. The kame deposits which occur above both sides of the gorge were probably deposited in hollows in thin and stagnant ice. Since the principal direction taken by melt-waters within the Vale of Stretton was towards the south, and since in the later stages of deglaciation the glacier drainage was appropriated along valley lines, most of the meltwaters must have passed through the Marshbrook gorge towards the Onny valley.

South of Marshbrook prominent terraces of fluvio-glacial sands and gravels extend on both sides of the Quinney brook and river Onny towards Ludlow and Brimfield. Watts found numerous erratics of Breidden hill dolerites in the gravels at Ludlow, and Dwerryhouse and Miller later found Eskdale granites and Ennerdale granophyres in the deposit.[1] Since these terraces may be traced into the Marshbrook gorge, it is likely that they are outwash deposits from the Church Stretton lobe of the Irish Sea glacier.

On the northern slopes of the Long Mynd the early stages of deglaciation were also characterized by the erosion of meltwater channels at relatively high elevations. The ice extended up to at least 1,300 ft. in this area, and in the west was confluent with the Welsh ice. As the ice gradually downwasted, the main watershed of the Long Mynd became exposed as a nunatak, and the Golden valley channel was eroded as a col gully. It was cut by meltwaters flowing through

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1. Dwerryhouse, A.R., & A.A. Miller, 1930, p. 121.

a shallow pass at the bipartation of the shrinking ice mass.

Subsequently, the Irish Sea and Welsh ice sheets became separated, and the latter gradually downwasted against the northern slopes of the Long Mynd. Most channels at high elevations head directly down the hillside, and it appears they were cut by meltwaters plunging directly into and beneath the ice margin. There was relatively little drainage parallel to the ice.

Some channels run for considerable distances down the hillside, though the morphology of several suggests that different portions were cut at different stages of deglaciation. For example, those channels, including Long Rough, which head northwards down Cothercott hill were initiated when the ice margin was at about 1,300 ft. At the foot of Cothercott hill they are tributary to a large channel heading north-east at the back of an erosion surface; the latter was cut submarginally when the ice surface was at about 900-1,000ft. Similarly, the Dead-mansbatch channel was initiated when the ice was at 1,300 ft. Lower downslope it grades into a pre-glacial valley on the side of which is a marginal channel and chute, and in the floor of which is incised another channel. The latter were probably formed when the ice had downwasted to about 900 ft.

Kames and other fluvio-glacial material also appear to have been deposited at a relatively late phase of deglaciation. There are none at high levels, all being confined to areas below 1,000 ft. Among the highest is the group of kames near Sheppen Fields and Underhill Hall, and those at Wildreck. There is a shallow kettle hole in one to the north of Underhill Hall, but no recognizable ice-contact slopes.

A 6 ft. section in one of the kames showed roughly-bedded and poorly-sorted sands and gravels, The kettle hole indicates they were deposited at the stagnant margin of the ice sheet. This is confirmed by the occurrence, adjacent to both kame areas, of the intakes of several meltwater channels all of which head directly downslope. The meltwaters must have flowed directly into the ice, instead of along the ice margin, suggesting it was relatively thin and stagnant.

Near Picklescott there are several ridges aligned parallel to the contours. They are composed predominantly of sands and gravels, and probably represent crevasse fillings.

On the lower slopes of the Long Mynd, between Woolstaston and Smethcott, is a complex area of fluvio-glacial deposition and erosion. A large number of kemes occur near Smethcott, many of which are plastered onto the sides of others. The topography around Woolstaston is rather more subdued, with extensive flat areas underlain by fluvio-glacial sands and gravels. There are numerous kettle holes scattered throughout the area. Borehole records prove that complex deposits underlie the area; sand, gravel, stony clay and clay alternate without any apparent regularity. A large meltwater channel cuts directly across the deposits; though bedrock is exposed in the upper portion, it is cut almost entirely into glacial deposits and truncates the sides of several kames. Chutes plunge down the southern side of the channel.

Whitehead interpreted these deposits as frontal moraines and morainic materials, with some sands and gravels, and consider that most of the drifts were deposited in a lake, the level of which fell progressively as the ice-front receded.[1] This view is not supported by

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1. Pocock, R.W., et. al., 152, 1938, p. 195.

the form and composition of the deposits. The large number of kettle holes and kames, and the general complexity of the topography, indicate that deposition took place in thin crevassed and stagnant ice. The meltwater channel was also eroded subglacially, probably slightly before the deposition of the kames. An analogous situation has been described by Niewiarowski in Poland where subglacial channels and kames are juxtasoposed.[1] The former were developed in the active ice phase and later, in the stagnant ice phase, they were filled with ice from the greezing of meltwaters (winter ice), while in crevasses and depressions in the stagnant ice kames were formed alongside, or on top of, the preserved channels. After the melting of the ice the channel would be exhumed and the kame deposits subside into it. This process may account for the truncation of several kames by the channel near Smethcott. On the other hand, both may have been formed together, within stagnant ice, with a subglacial channel cutting into several kames after they were deposited. The subglacial chutes on the southern side of the channel were probably cut by meltwaters plunging into dead ice within the channel at a fairly late stage of deglaciation.

To the north of Smethcott is a smaller area of fluvio-glacial deposition, with kames standing above a surface of ground moraine. Some are isolated, and in the flanks of several are kettle holes up to 15 ft. deep, indicating the former presence of dead ice.

Several phases of deglaciation may be recognized on the north slopes of the Long Mynd. The first was characterised by the cutting of

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1. Niewiarowski, W., 1965, p. 11-12.

meltwater channels at a high level, several having intakes at 1,300 ft. A second phase occurred when the ice had downwasted to about 900 ft. Many new channels were cut, and kames were deposited within the stagnant ice margin. Subsequently, as the ice receded still further, the complex kame and kettle topography between Smethcott and Woolstaston was deposited.

On leaving the Long Mynd the ice retreated northwards towards the Lyth hill and Sharpstone hill ridge. To the south of the former, on both sides of Cound Brook, is the prominent Dorrington-Ryton-Stapleton esker system. Various theories have been proposed to explain these eskers. Barber[1] believed they are probably the deltaic deposits of glacial rivers, formed where such streams emptied into a lake at the ice front. This lake was impounded between the Long Mynd, Caer Caradoc and Lawley to the south, and the front of the ice sheet to the north, and presumably also extended into the Vale of Stretton. Whitehead[2], on the other hand, doubted whether the ridges could be regarded as eskers. He thought that they were either kames, deposited at an ice margin aligned NW-SE; or were part of a once-continuous spread forming part of a large kettle moraine left by the melting of a stagnant lobe of ice when the main ice front had retreated beyond the Lyth Hill and Sharpstone hill ridge. In other words, the eskers were thought to be primarily erosional features.

The origin of the esker system is important in determining the condition of the ice at the time of formation, and whether or not any

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1. Barber, J., 1922, p. 31.
2. Pocock, R.W., et. al., 152, 1938, p. 193.

proglacial lake existed. It seems highly unlikely that it is of erosional origin. There is no direct connection with the kame complex around Bomere, from which it is separated by a wide expanse of ground moraine. The eskers comprise a series of sinuous ridges which join the subdivide just like the distributory pattern of a glacial stream, and this would favour a depositional, rather than an erosional, origin. Also, the esker system cannot be regarded as a series of kames. A kame is essentially,

> " ... a mound composed chiefly of sand and gravel ..."[1],

and may have some affinity with eskers in that a thin band of deposits may link the mounds. This cannot be reconcilèd with the form of this feature since it is not beaded and does not comprise a series of mounds joined together, but is a group of sinuous ridges with even surfaces.

Esker deposits may be laid down, as envisaged by Barber, by sub-glacial or englacial meltwaters as they emerge from a glacier into ponded water. However, this would lead to the formation of a beaded esker as the ice front receded. Short narrow ridges would be separated by deltaic fans, built where the stream was free to spread beyond the confining walls of a tunnel. Again, this origin is not consistent with the form of the esker system. Further, the prevalence of stagnant ice features to the south of the esker system argues against the existence of a large pro-glacial lake. With downwasting playing such an important part in deglaciation, it is unlikely that any significant amount of fluvio-glacial deposition took place proglacially.

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1. Holmes, C.D., 1947, p. 248.

The form of the esker system and its internal composition suggest that it originated subglacially; the material was deposited in tunnels at the base of the glacier when the ice was thin and stagnant, or nearly so. The sinuous ridges closely resemble a distributory stream pattern, and the cross-bedding of the sands and gravels represent stream deposits. The occurrence of a kettle hole between two of the ridges and in the flank of another is indicative of dead ice. The fact that the ridges attain the same general level indicate that their upward limit was controlled by some agent common to both, possibly an englacial water-table. Subglacial eskers usually trend at right angles to the ice front, which must in this case have been aligned ENE-WSW, and not NW-SE as proposed by Whitehead. Glacial striae often record the latest direction of ice flow; some on Sharpstone hill are aligned roughly north-south, and confirm the alignment of the ice front deduced from the direction of the eskers.

It would appear that a large mass of stagnant ice existed in front of the Bayston hill and Sharpstone hill ridge, and this was probably continuous with a similar spread in which the kame complex around Bomere and Berrington was deposited.

## DEGLACIATION OF THE WELSH ICE SHEET

Referring to deglaciation to the west of the Long Mynd, Whitehead stated that,

> "In the valley of the Habberley Brook...
> a lake could hardly have failed to be

> produced by ice, blocking its northern exit."[1]

Evidence for this lake included overflow channels on the side of the valley and their relationship with fluvio-glacial gravels on the valley floor. The ggravels at Pollardine were believed to have been deposited in a lake that had an outlet across the Stiperstones through the Eastridge channel, at 1,000 ft. The surface of the gravels at Marsley is at 700 ft., and these were supposedly deposited by a torrent from the Eastridge valley in a lake that had fallen to nearly this level. The lower lake is believed to have had two outlets, both of which functioned at different times. These were the Granham's Moor channel, whose floor is at 800 ft., and the channel at Broomhill on the eastern side of the valley. The gravels at Habberley were believed to be either morainic, or the remains of a delta, formed when the lake was at a still lower level.

It has been demonstrated earlier that the former existence of a glacial lake in the Habberley valley cannot be substantiated from the form and disposition of the meltwater channels. None are lake overflows, and both the Eastridge channel and the one at Granham's Moor carried meltwaters into, and not out of, the valley; both were eroded sub-marginally. Other channels in the area suggest a gradual downwastage of ice, and it is in this light that the origin of the gravel deposits on the valley floor must be considered.

The gravels at Pollardine, the surface of which is at 1025 ft.

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1. Pockock et. al., 152, 1938, p. 203.

form a kame deposit; they are unsorted and poorly-bedded, which is inconsistent with deposition in a lake. Other, smaller kames occur nearby. The deposit at Marsely is a relatively flat-topped mound, the surface of which slopes down towards the centre of the valley. Its position opposite the mouth of the Eastridge channel suggests it was laid down by meltwaters issuing from that valley. However, the kettle hole on the north-eastern side of the mound indicates deposition is associated with stagnant ice, rather than in a glacial lake. The steep edge on the north could be an ice-contact slope. The mound is composed of poorly sorted, subrounded gravels at the surface, though the internal composition is not known. The gravels are mainly of Stiperstones quartzites, which also suggests it was deposited by melt-waters which cut the Eastridge channel. The mound is interpreted as a fluvio-glacial fan deposited against, or within, a stagnant ice margin. It was probably formed at a later stage of deglaciation than the kame at Pollardine, when the ice had downwasted to about 750 ft.

To the west of Habberley village a conspicuous terrace of fluvio-glacial sands and gravels has been laid down against the Stiperstones, immediately to the south of the mouth of the Granham's Moor channel. The terrace has a steep western face, probably an ice-contact slope. The feature is a kame terrace deposited against the margin of the ice sheet when its surface was at about 750 ft. Since the Granham's Moor channel is of subglacial origin, much of the material forming the terrace may have been derived from the channel and deposited beneath the ice margin in a small lake. The terrace has subsequently been cut into at its northern edge by meltwaters from the channel.

North of Habberley, the irregular fluvio-glacial hillocks are interpreted as kame deposits, laid down when the ice had downwasted still further. In this sense they are morainic, but they are not the remains of a delta. They do nott have deltaic form and they show no signs of subsequent erosion.

At the glacial maximum ice filled the Habberley valley, reaching up to at least 1,400 ft. against the Stiperstones, the summit of which may have protruded above the ice as a nunatak. During deglaciation the ice gradually downwasted and retreated. The Stiperstones and the Longmyndian hills bordering the valley were progressively uncovered. Meltwater channels were eroded at high levels, the intakes of those on the Stiperstones being at a maximum of 1,400 ft. When the ice stood at about 1,100 ft. the Eastridge channel was initiated; it was cut into the floor of a pre-glacial valley and carried meltwaters eastwards into the ice in the Habberley valley. The marginal channel west of Pollardine was also eroded at about this time, and the nearby kame deposits laid down. The channel at Granham's Moor was eroded sub-glacially when the ice surface had downwasted to about 800 ft., and it also carried meltwaters into the Habberley valley. The fan at Marsely and the kame terrace at Habberley were deposited at this stage. Also, when the ice was at this level, the channel between Lawn Hill and Bank was cut, directing meltwaters across the Longmyndian hills to the east.

The slope and direction of the meltwater channels gives an indication of the slope of the ice surface during deglaciation. When the ice was at 800 ft. meltwaters were flowing generally eastwards, both into and

out of the valley, from the Welsh into the Irish Sea ice. The ice surface must, therefore have sloped down in this direction. With a further lowering of the ice surface there was no longer any direct escape route for the meltwaters across the Longmyndian hills. It was at this stage that the gorge to the east of Pontesford and Earl's hills was cut, by the superimposition of an englacial stream across a divide separating two pre-existing valleys. Initially, the meltwaters flowed north-eastwards through the col at Radlith, and cut the sub-glacial channels at Lingcroft Pool. This tendency for the meltwaters to flow eastwards also reflects the general eastward slope of the ice surface. With later incision and downwasting the meltwaters flowed northwards towards Pontesbury.

The Shelve district, to the west of the Stiperstones, underwent a similar pattern of deglaciation. A smaller proportion of fluvio-glacial material was deposited, and the evidence lies mainly in the form and disposition of the channels. Submarginal and subglacial drainage was predominant, with meltwaters heading both north and south away from a low divide passing close to Shelve village. Of the southern channels, that at Ritton Castle is the largest; it was eroded sub-marginally to the Brook Hill ridge. No large channels are discordant to the topography, and the meltwaters were chiefly concentrated in pre-existing valley lines. As Corndon Hill was uncovered a system of submarginal drainage was initiated on its flanks, the meltwaters heading generally eastwards. With the thinning and retreat of the ice margin a few kames were deposited in the Black Marsh and Hyssington Marsh areas. These may indicate successive positions of the ice front in

in the upper West Onny valley.

The largest channel which heads northwards away from Shelve is the Hope valley. The morphology of the channel indicates that it was eroded in two separate phases of deglaciation. The upper part of the channel, above Hope, was incised submarginally into a broad bench on the side of a narrow volcanic ridge. Numerous chutes on both sides of the channel indicate that it was filled with stagnant ice at a later phase. The section of the channel below Hope was cut after a significant retreat of the ice margin. Englacial meltwaters were incised across the volcanic ridge, cutting into the floor of the previous channel, and then ran along the other side of the ridge towards Ploxgreen. Chutes on the southern side of the channel suggest a sub-marginal origin. Kames and other fluvio-glacial material were deposited around Snailbeach and Wagbeach, and probably indicate the margin of the ice sheet at a later phase of deglaciation.

Deglaciation in the other upland areas followed a similar pattern to that in the Long Mynd-Stiperstones-Shelve district. There was a gradual downwasting and stagnating of the ice sheet until it eventually became confined to the broad valleys separating the hills. There is considerably less fluvio-glacial material associated with the Welsh ice, and the hills are more or less devoid of kames and similar forms. The principal evidence for deglaciation in these areas lies in the melt-water channels.

The Long Mountain was completely covered by ice at the glacial maximum, and meltwater channels are found right up to the main water-

shed, at 1,275 ft. They are located mainly on the southeast and eastern side of the mountain. In the early stages of deglaciation englacial meltwaters were superimposed onto the broad valley heads, and thereafter the meltwaters were confined to the pre-existing valleys. Several have submarginal channels on their sides, while the floors and sides of many have been considerably overdeepened and steepened. The pre-existing topography guided much of the meltwater drainage on the Long Mountain. There are few channels which are cut across spurs or watersheds: these are confined to the highest levels and were formed by a superimposition of englacial or subglacial drainage.

On the Kerry hills the principal area of fluvio-glacial erosion is concentrated on the hillslopes above Sarn. Here, a system of submarginal and subglacial channels is found up to 1,400 ft. The early phases of deglaciation were characterised by the erosion of submarginal channels at high levels. One of these was superimposed into the floor of a col across the main watershed, and directed meltwaters into the headwaters of the river Unk. Other, smaller, submarginal channels also occur on the hillslopes to the west of Sarn. Later, with ice thinning and stagnation, the submarginal drainage was replaced by meltwaters plunging directly into the ice, cutting a series of subglacial chutes.

The direction of slope of the submarginal channels indicates the slope of the ice surface. On the northern side of the Kerry Hills, all the channels slope down towards the east, indicating that the slope of the ice in the Mule-Caebitra valley was in this direction.

Several submarginal channels also occur on the Kerry Hills overlooking the Camlad valley. These are at a lower level than those above Sarn, lying between 800 and 700 ft. They are succeeded by chutes, formed at a later stage by meltwaters plunging directly into the ice. These submarginal channels were probably formed after those above Sarn, when the ice surface had downwasted to about 800 ft. They slope down towards the east, and the meltwaters were being directed towards the Plowden gorge at the southern end of the Long Mynd.

Ultimately, the ice downwasted from the hill areas and became confined within the limits of the broad valleys. The final stage of deglaciation formed an image of the initial stage of glaciation. The ice sheet broke up into a series of valley glaciers which gradually retreated towards the Welsh uplands. Although it is difficult to detect definite stages, the progress of deglaciation is seen more clearly in the valleys than it was on the hills. The ice was considerably more stagnant than when it lay over the hills, and consequently there was a greater volume of meltwaters drainage and of fluvio-glacial deposition.

A mass of stagnant ice occupied the Plowden gap at the southern end of the Long Mynd, beneath which a large number of channels was eroded. The largest of these is the Plowden gorge, through which flows the river Onny. It has been incised some 25 ft. below a broad 600 ft. platform within the gap, and may be traced continuously from Plowden in the west to Wistanstow in the east. Chutes on both sides of the gorge testify to a subglacial origin. The dimensions of the gorge indicate that it carried a large volume of glacial meltwaters. It

formed the principal outlet for meltwaters from the upper Onny basin.

Small fluvio-glacial terraces occur on both sides of the gorge, and beyond its exit they may be traced, at different levels, to beyond Wistanstow where they join those of the Quinney Brook. The latter are outwash terraces deposited by meltwaters from the Irish Sea ice flowing through the Marshbrook gorge. Those within the Plowden gap are also outwash terraced, but in this case are related to the melting of the Welsh ice sheet. The proglacial meltwater streams of both ice sheets were confluent at Wistanstow, and continued southwards down the Onny valley towards Craven Arms and Ludlow.

A fine series of marginal meltwater channels associated with ice in the upper Onny basin are located on the hillsides around Norbury and Wentnor. Above Norbury, three channels are in direct line with one another; they are cut across spurs and one has a humped long profile. They slope down towards the east at an angle of about $3^{o}$ and eventually plunge down the side of the East Onny valley. These are submarginal channels, and were probably formed when the ice surface was at about 900 ft. Other, similar, channels are found at lower elevations, and are probably related to a slightly later phase of deglaciation, when the ice surface had downwasted still further.

Immediately south of Wentnor, two channels cut across a prominent ridge at 750 ft. Since they both have humped long profiles, and chutes plunge down their sides, they were probably cut submarginally.

On the opposite side of the Onny basin a series of channels head obliquely and directly down the hillside towards the river. Their intakes are between 750 and 600 ft., and they were eroded in a sub-

marginal position.

These channels were all formed marginally or submarginally to a mass of ice within the Onny basin. They all slope down towards the east, concentrating the meltwaters towards the Plowden gorge.

Extending in a large arc across the Onny basin from Bishop's Castle, through Lyndham Head and More and a short way up the Camlad valley is a region of hummocky moraine topography. This is not a till moraine, but comprises a complex of kame deposits, ridges and numerous kettle holes. It forms part of Charlesworth's South Wales End Moraine[1], but is clearly a recessional rather than a terminal feature. Smaller kame complexes occur in front of the moraine at Hardwick and Eaton, and there is a small kame to the north of the gorge at Plowden. The presence of kettle holes within this complex area of fluvio-glacial deposition indicates that it was formed at a stagnant ice margin.

During deglaciation the main body of the ice will remain dynamically active, even during quite rapid retreat, but the ice commonly has stagnant zone at the frontal margin and the sides. Stagnant ice margins are fairly common in present-day temperate glaciers in Alaska and Spitsbergen. Unlike the relatively clear active ice behind, the stagnant margins often have a thick cover of glacial and fluvio-glacial debris, in addition to extensive subglacial and englacial fluvial deposits. Kame moraines, also known as hummocky moraine, or dead-ice moraine, have been observed in the process of formation at the front of many glaciers.

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1. Charlesworth, J.K., 1929, p. 349.

Lamplugh[1] and Boulton[2] studied them at the margins of glaciers in Spitsbergen, and the latter showed that they may be developed sub-aerially after sediment has been deposited on the glacier surface from a series of englacial debris bands. The debris is mainly fluvio-glacial material picked up from the glacier bed and transported along planar structures to the surface. The material is deposited over buried ice, itself full of fluvial sediment and abundant debris, and on final melting gives rise to a complex hummocky kettle moraine.

Hoppe[3] holds different views with regard to the origin of hummocky moraine in Sweden. These are formed mainly of till, and characteristic features include ridges, dead-ice hollows and moraine plateaux. Following the work of Grigg[4] in Spitsbergen, Hoppe believes that crevasses form in stagnant ice and subglacial till, soaked by meltwater, forms a plastic mass which is squeezed up into basal cavities due to the load pressure of the ice on its substratum.

The lack of till in the Lydham Heath moraine, the predominance of fluvio-glacial material, the complex dead-ice topography, points to an origin similar to that outlined by Boulton. The ice in this terminal area would be covered by a concentration of fluvio-glacial debris, brought to the surface along shear planes. As the buried ice melted away, sliding and slumping of the debris would occur, leaving a maze of mounds and kettle holes.

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1. Lamplugh, G.W., 1911.
2. Boulton, G.W., 1967.
3. Hoppe, G., 1952.
4. Grigg, K., 1929.

In front of the moraine is a large spread of gravels which may be traced along the river Onny into the Plowden gorge. These are outwash gravels, deposited proglacially by meltwaters flowing from the stagnant ice margin at Lydham Heath. The role of the Plowden gorge as an escape route for glacial meltwaters from ice within the upper Onny basin is again emphasised.

Between Lydham Heath and Church Stoke there are no dead-ice features, indicating a fairly rapid retreat of the ice. There are, however, several meltwater channels on the valley sides. At Hyssington a complex channel system was eroded by the superimposition of englacial meltwaters. They head generally southeastwards, towards the Camlad valley. To the south, at Upper Snead, a series of submarginal and subglacial channels run down the hillside. These channels, and those at Hyssington, were formed at a stagnant ice margin. On the opposite side of the Camlad valley are a number of submarginal channels and chutes. The former occur between 800 and 700 ft., and decline towards the east.

A complex group of deglaciation landforms occur in the neighbourhood of Church Stoke. The most prominent of these is the gorge of Marrington Dingle. The nature and origin of the gorge has been discussed in some detail in Chapter 5. Both Harmer[1] and Dwerryhouse and Miller[2] believe it operated as an overflow channel from a proglacial

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1. Harmer, F.W., 1907, p. 470.
2. Dwerryhouse, A.R. & A.A. Miller, 1930, p. 102.

lake occupying the depression south of Church Stoke. The waters supposedly flowed along the edge of a lobe of ice within the Montgomery-Chirbury gap.

> " ... the gorge was cut partly by the surplus waters of a small lake ponded up between the ice front and the hills, but chiefly by water from melting ice."[1]

It has been demonstrated that the gorge did not originate as a glacial lake overflow channel. It was formed essentially in a subglacial environment, by the superimposition of englacial meltwaters onto the underlying topography. Small chutes and marginal channels are found on both sides of the gorge, the outlets of some of which hang well above its floor. The southern half of the channel occupies the floor of a pre-existing valley which drained in the opposite direction. The channel cuts across a prominent watershed, and there has been a reversal of the pre-glacial drainage. The direction of the meltwater flow must have been determined by the slope of the ice surface, rather than the underlying topography. Thus at the time the gorge was cut the ice gradient in the Montgomery-Chirbury gap must have been down towards the north. This ice was contiguous with that in the Caebitra-Camlad valley; it has been shown that the ice in the upper Camlad valley sloped eastwards and that meltwaters were directed towards the Plowden gorge. Marrington Dingle must have functioned in a similar manner to the latter, carrying waters from the Caebitra valley across the Montgomery-Chirbury gap into the Marton valley.

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1. Ibid.

A prominent drift terrace extends across the Marton valley immediately upstream from the exit of Marrington Dingle. The terrace now forms a watershed in the valley floor, and the Camlad is prevented from flowing northeastwards to join the Severn in the plain of Shropshire. Instead, it turns sharply to the south-west, and enters the Severn near Forden. The terrace is composed cheifly of fluvioglacial sands and gravels, though exposures in its steep lobate edge show some till. The gravels extend northeastwards down the valley, and continue beneath peat deposits adjacent to Marton Pool. The surface of the terrace is pitted with kettle holes, and Marton Pool is considered to be a particularly large example. At the northern edge of the terrace, close to Stockton, are several kames.

The terrace is essentially an outwash deposit, the kames and the steep lobate edge represent the ice margin. The kettle holes were formed in the melting of detached masses of dead ice in the outwash. Since the outlet of Marrington Dingle cuts partly into the terrace it must have been eroded either during, or after, its deposition. The former seems most likely, and the meltwaters from the gorge would flow into stagnant ice in the Marton valley. Material derived from the excavation of Marrington Dingle would contribute to the construction of the outwash terrace. The meltwaters from the gorge would join those of the stagnant ice margin and flow northeastwards down the Marton valley as outwash. After the regreat of the ice, the terrace acted as a barrier to the Camlad, which was compelled to turn southwestwards.

Dwerryhouse and Miller[1] make certain deductions regarding glacial chronology from the relationships between Marrington Dingle and the supposed glacial lake Lapworth. They infer that the Marton valley, east of Hockleton Bridge, must have been free from ice, to allow the escape of the water from Marrington Dingle.

> "The level of the lake on the plain of Shropshire must at this stage, have been below 335 ft. O.D ..."[2],

the level of the valley floor at Marton Pool.

> " ... Marrington Dingle must belong to the later stage when Glacial Lake Lapworth (300 ft. O.D.) was in existence."[3]

It is not doubted that the Marton valley east of Hodkleton Bridge was free from ice when Marrington Dingle was cut, but there is no evidence to suggest that the meltwaters flowed into an extension of glacial lake Lapworth. The conclusion that

> " ... the period of the maximum of this portion of the Welsh ice was subsequent to the greatest extension of the Irish Sea glacier on the Shropshire Plain"[4],

is not valid. The fluvio-glacial terrace does not mark the maximum extension of the Welsh ice in this area, but merely represents a retreat phase of the ice sheet.

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1. Op. cit., p. 102-103.
2. Op. cit., p. 103.
3. Ibid.
4. Ibid.

At the southern edge of the Montgomery-Chirbury gap, in the grounds of Llymore Park and extending towards Rhishton, are several kames which were laid down marginal to ice within the Caebitra valley. Much of the gap was probably free from ice at this time. Similar kame deposits are found on the northern edge of the gap, and these were formed marginal to ice within the lower Camlad valley, possibly after it had retreated from Hockleton Bridge. The kame deposits represent relatively late stages of deglaciation in this area.

West of Church Stoke, the broad flat floor of the Caebitra valley is mostly covered by extensive spreads of well-bedded gravels. They may be traced upstream on both sides of the valley towards Sarn. Immediately to the west of Church Stoke the gravels are pitted by numerous shallow kettle holes.

Dwerryhouse and Miller[1] interpret the gravels as the products of torrents from the ice, the front of which had retreated to Pentrehyling or beyond, into the waters of a lake which occupied the triangular depression south of Church Stoke. The overflow waters of the lake are those which cut Marrington Dingle. There is, however, no evidence of a glacial lake in this area. The numerous kettle holes near Church Stoke were formed on the melting of detached masses of dead ice within the gravels, outwash material deposited in front of the receding glacier. The bedded brick-earth at Church Stoke, mentioned by Harmer[2], is not a lake deposit, but fine outwash sediment laid down some distance in

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1. Op. cit., p. 118.
2. Harmer, F.W., p. 481.

in front of the ice margin.

To the west of Sarn, a further pause in the retreat of the Welsh ice is marked by the kame moraine which extends in an arc across the Mule valley. The moraine comprises a complex of kames and kettle holes, and was formed at the stagnant ice margin. It forms a watershed, at 600 ft., in the Mule-Caebitra valley. The pre-glacial course of the Mule was by the open valley, an obvious continuation of the part between Kerry and Glan Mule, past Sarn and down the wise Caebitra valley towards Church Stoke.[1] Immediately to the west of the moraine, the Mule now makes a sharp turn towards the north-west, and flows into the Severn by way of the Mule gorge.

Between the Sarn moraine and Upper Maenllwyd the Mule flows in a broad drift-filled valley. It then enters the deep, steep-sided gorge, which cuts directly across the relief, more or less at right angles, truncates several spurs and ridges and is markedly asymmetrical in cross-profile. It was formed by the superimposition of englacial meltwaters onto the underlying topography. The gorge entrance is at about 500 ft., and its higher, eastern, side about 600 ft. The height of the moraine at Sarn is also about 600 ft., and it is probable that the gorge was initiated when the ice-front stood in this position in the Mule-Caebitra valley. The ice lying over the hills to the north would be relatively thin, and glacial meltwaters would tend to flow towards the Severn valley, which is at about 300 ft., rather than north-eastwards across the Cefn-y-Coed hills. There is no pro-glacial

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1. The pre-glacial drainage and subsequent drainage derangement will be discussed in the next Chapter.

outwash between the moraine front and Sarn village, suggesting that, when the ice was in this position its meltwaters flowed not towards Church Stoke and Marrington Dingle, but took a shorter and easier route through the Mule gorge to the Severn valley. Upon the retreat of the ice, the moraine acted as a permanent watershed, and the Mule is now diverted through the gorge to the Severn.

Upstream of the Sarn moraine there are no major deglaciation landforms in the Mule valley. Well-bedded outwash sands and gravels are spread over the valley floor towards Kerry, and may be traced some way up the tributary Miheli brook. These were deposited by pro-glacial meltwaters from a relatively rapidly retreating ice front. It is difficult to assess the thickness of the outwash deposits, but up to 10 ft. were observed in river bank sections. Over 100 ft. of glacial and fluvio-glacial deposits are proved by boreholes at both Kerry and Pen-y-gelli.

There is no evidence that the Welsh ice became detatched and stagnanted *in situ* in the Mule valley. The ice front apparently retreated southwards up the major valleys draining the Kerry Hills, particularly the Mule and Miheli.

In the upper part of the Miheli valley is a group of flat-topped terraces, composed of glacial and fluvio-glacial material and extending downstream as a convex lobe. This is not a true arcuate terminal moraine, but was probably deposited during the deglaciation phase. The fact that till in places overlies sands and gravels indicates a slight re-advance of ice in this area. The deposit has been dissected by glacial melt-

waters, for example at Middle-Cwm-y-ddalfa.

There are a number of glacial meltwater channels in the headwaters of the Mule valley. The hillsides around Wig and Cwm-y-gigfran are dissected by submarginal and subglacial channels, formed by meltwaters plunging beneath stagnant ice.

At the head of the Mule, the moraine in Black Gate col is dissected by several meltwater channels, probably formed when the ice front lay close to the watershed. Other areas of the watershed to the south have also been modified by fluvio-glacial erosion and deposition.[1] Dwerry-house and Miller refer to a well-marked dry channel which was

> " ... formed by the drainage from a lobe of ice in the Ithon valley, at a period of its retreat when it was no longer able to surmound the watershed."[2]

Subsequently, the ice fell below the level of the watershed and continued its retreat towards the mountains of Central Wales.

In the Severn and Marton valleys the retreat of the ice was relatively rapid and, except for the feature at Marton, there are no recessional moraines. There are a few isolated kames in the Marton valley to the south-west of Asterley, but elsewhere the floor is masked by peat and alluvium, overlying outwash gravels laid down in front of the retreating ice. The hillsides bordering the valley are dissected by meltwater channels, including many of submarginal and subglacial origin.

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1. Luckman, B.H., 1966, p. 80-100.
2. Dwerryhouse, A.R. & A.A. Miller, 1930, p. 100.

There is no evidence that a glacial lake ever existed in the valley.

Dwerryhouse and Miller maintained that,

> "The valley of the Severn between Newtown and Welshpool shows the usual features left by a retreating glacier up a main valley, namely small frontal moraines athwart the stream, each having an alluvial fan above it. A good example of such a moraine occurs near Llegodig, about 5 miles below Newtown."[1]

These supposed moraines have not been found in the Severn valley. The mounds in the valley floor near Llegodig are bedrock features, moulded into streamline form by the passage of ice over them.[2] They are similar in form and origin to other mounds about a mile downstream at Court Calmore.

Between the Breidden Hills and Berriew the valley floor is relatively featureless and flat; the river meanders over a broad alluvial flood plain which overlies well-bedded gravels, probably outwash material. There are several low terraces, particularly where tributary valleys enter the Severn. Two gravel terraces, at 5 and 10ft. above the river, extend into the Severn valley from the mouth of the Rhiw. These are probably post-glacial features and not related to deglaciation.

At Abermule, at the mouth of the Mule gorge, there is a considerable thickness of glacial drift and several low terraces. Borehole records

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1. Dwerryhouse, A.R., & A.A. Miller, 1930, p. 99.
2. See Chapter 2. p. 65

show the drift to be thickest on the southeastern side of the valley, while on the opposite side the river is entrenched in bedrock. The drift mainly comprises marls, sands and gravels, and it is likely that they were laid down as a fan, by meltwaters issuing from the Mule gorge. Low terraces, 5-10 ft. high border the Mule as it crosses the Severn valley, and there is a 15 ft. terrace on the left bank of the valley. Again, these are probably post-glacial features.

The right flank of the Severn valley between Newtown and the junction of the Mule is deeply covered with drift. The river itself is often entrenched into bedrock on the opposite side of the valley. There are several drift terraces, rising up to 35 or 40 ft. above the river, and at Aber Bechan is an area of hummocky morainic topography. Kame mounds stand on the terraces, and lower terraces extend downstream from them. The drift in this part of the Severn is principally outwash gravels, and the kames mark minor pauses in the retreat of the ice. Other prominent kames occur above Newtown and have outwash terraces leading away from them.

Apart from the outwash gravels and occasional kames, there are few deglaciation landforms in the Severn valley. The ice underwent a fairly rapid retreat towards Central Wales.

## CONCLUSIONS

The landforms of fluvio-glacial erosion and deposition indicate that the deglaciation of the Welsh and Irish Sea ice sheets took the form of gradual downwastage from the hills, followed by retreat up the main valleys towards the high mountains of Central Wales on the one hand,

and the Shropshire-Cheshire basin on the other.

Numerous meltwater channels show that there was a considerable volume of subglacial and englacial drainage, and that the margins of the ice were largely stagnant. In the higher areas the channels are often discordant and have humped long profiles; they were formed by the superimposition of englacial meltwaters onto the underlying relief. Lower down, with continued ice thinning, the topographical control increased and the meltwaters became concentrated in valleys. Several prominent gorges were eroded, often cutting across pre-glacial watersheds.

When the ice had uncovered the hills and was confined to the broad valleys, the gorges formed the principal outlets for the pro-glacial outwash, each operating at different phases of deglaciation. The Marshbrook gorge was the main outlet for meltwaters from the Irish Sea ice. Outwash terraces were deposited on both sides of the Quinney brook and river Onny, and may be traced beyond Craven Arms to Ludlow and Brimfield. The Plowden gorge was the outlet for meltwaters from the Onny basin, also directing them towards Craven Arms and Ludlow. With the retreat of the Welsh ice to Church Stoke, the Plowden gorge was replaced by Marrington Dingle as the major outlet. Meltwaters were now directed, via the Marton valley, towards the Severn. The further retreat of the ice to the Sarn moraine initiated the Mule gorge as the outlet, and the meltwaters took a shorter and easier route to the Severn.

Pauses in the retreat of the ice are marked by considerable accumulations of kame and kettle moraine. They were deposited within

a stagnant and dead ice margin, though the latter did not necessarily become detatched from the active ice behind. Finally, the ice retreated towards the high mountains of Central Wales, and across the Shropshire-Cheshire plain towards the Irish Sea basin.

The classical view of deglaciation is incorrect in this area. Earlier workers had postulated the formation of large pro-glacial lakes and the cutting of the main gorges as overflow channels. Nowhere is there evidence that such lakes existed, and the gorges were formed by the superimposition of englacial meltwaters onto the underlying relief. In particular, the supposed lake in the Vale of Stretton and part of the plain of Shropshire is not consistent with the landforms of deglaciation. Downwasting and stagnating of the ice occurred in these areas. This must cast doubt on the existence of Lake Lapworth as envisaged by Wills[1], and later by Poole and Whiteman.[2] While the former showed the lake as a small glacially-impounded feature occupying the area east and south of Shrewsbury, the latter believe that the 300 ft. lake in fact covered the whole of the south and eastern part of the Shropshire-Cheshire basin and extended north into the Wigan district. A higher lake level, at 330 ft., is well-developed throughout the basin;

> " ... near Welshpool it forms a good feature first described by Dwerry-house and Miller,"[3]

The 'good feature' refers to Marrington Dingle and the Marton moraine.

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1. Wills, L.J.
2. Poole, E.G., & A.J. Whiteman, 1962, p. 113-118.
3. Op. cit., p. 116.

Both, however, are unrelated to any glacial lake, and the moraine was formed at a stagnant ice margin; it is decidedly not a shoreline feature. Small proglacial lakes may well have existed during deglaciation, but no large features were impounded.

Deglaciation was characterised by downwastage and retreat of the ice-sheets, interrupted by several pauses. There is no evidence of a major re-advance. The principal moraines are formed of fluvio-glacial material; not till as would be expected if they were re-advance features. In very few placed does till overlie sands and gravels, and nowhere does it overlie outwash.

# Chapter 8

# DRAINAGE EVOLUTION

> "Deciphering the drainage evolution in the Border country is an extremely complex task. There have been a large number of diversions by ice, and the present pattern is very different from that of pre-Pleistocene times." (E.H. Brown, 1960, p. 138-139)

## PRESENT DRAINAGE PATTERN

The present drainage pattern is extremely complicated and several rivers have quite remarkable courses (fig. 8.1). The principal rivers flow in broad vales separating the main hill masses. Low watersheds are located in the centre of each of the vales, and the streams flow from them in opposite directions. The valleys are filled with considerable accumulations of drift, and the watersheds are usually formed by morainic accumulations. Most rivers have a gorge section, a meltwater channel, which forms an interconnecting link between the vales.

The West Onny flows southwards from its source near Shelve, and then turns sharply towards the east at Lydham

where a kame moraine forms a watershed at 600 ft in the Camlad-Onny depression. At the southern end of the Long Mynd the Onny flows through the Plowden gorge towards Craven Arms.

The Onny is joined from the north by the Quinney Brook which drains the upper part of the Vale of Stretton through the Marshbrook gorge, also a meltwater channel. Drift forms a watershed in the Vale of Stretton, north of which the Cound Brook flows towards the Severn.

The Camlad rises in the Corndon hill area, immediately to the south of a prominent bend in the West Omny, and flows towards Lydham where it makes a sharp turn towards the west. At Church Stoke it meets the eastward-flowing Caebitra and the combined waters flow north through Marrington Dingle, across the Montgomery-Chirbury gap. At the mouth of the dingle the Camlad makes a sharp turn to the east and joins the Severn near Forden.

Between the Long Mountain and Shelve is a broad vale with a low watershed at Marton Pool. The Rea Brook drains north-eastwards to join the Severn at Shrewsbury, and the Aylesford Brook flows south-westwards into the Camlad, and hence into the Severn.

The Mule rises in Black Gate col on the Kerry hills

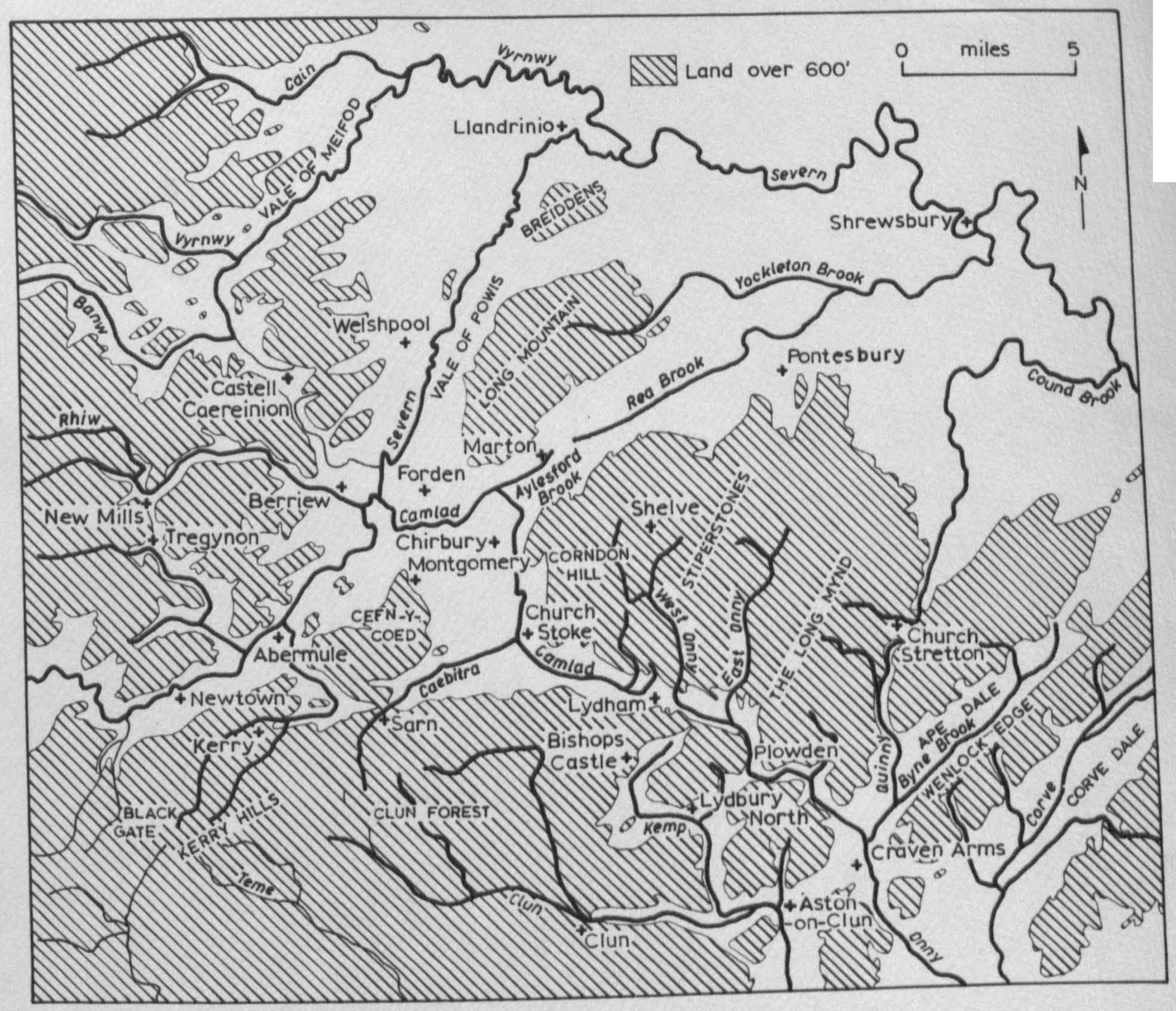


Fig 8.1 Drainage pattern

and enters a broad vale at Kerry, from whence it heads eastwards towards Sarn. A kame moraine deposited across the valley forms a watershed at 600 ft between the Mule and the Caebitra, and the Mule turns sharply towards the north and flows through the Mule gorge to join the Severn at Abermule.

Below Newtown there are several lowland courses open to the Severn and the Shropshire Plain; either north-eastwards along the Rea Brook-Camlad depression, or through the broad gap between the Long Mountain and the Breiddens. Instead, it takes the shortest route, along the Vale of Powis, and opens out onto the Shropshire Plain at Llandrinio. The Severn then turns eastwards and is joined by the Vyrnwy, and pursues a meandering course towards the Iron Bridge gorge. In pre-glacial times it is thought the Severn continued northwards beyond Llandrinio towards the Cheshire Basin, and was only diverted through the Iron Bridge gorge in late Pleistocene times.

## EARLIER THEORIES REGARDING PRE-GLACIAL DRAINAGE

One of the earliest attempts at elucidating the diversions of drainage in this area was made by Cobbold[1]

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1. Cobbold, E.S., 1926, p. 230-234

who believed them to be of glacial origin. A pre-glacial watershed supposedly existed at Marshbrook, and the drainage of the country around Marshbrook and Little Stretton was directed northwards. The Marshbrook gorge was cut as an overflow channel to a lake impounded in the Vale of Stretton. The diversion of the drainage remained permanent after the draining of the lake and the withdrawal of the ice. Cobbold also believed that the Plowden gorge is an overflow channel of a lake impounded in the Lydham-Church Stoke area,

> "...which was blocked off from its natural outlet to the Severn by ice, and had to rise to an altitude of perhaps 800 or 900 feet to escape over a col of Tarranon Shale at Plowden, where this formation is in contact with hard Longmyndian rocks - for such an altitude the low channel about Lydbury North would have to be blocked by ice (or moraine, of which there is no record known), at least 200 ft in thickness."[1]

Cobbold's views have, until recently, gained a general acceptance, and have been elaborated by Dwerryhouse and Miller[2] who reconstructed the pre-glacial drainage (fig. 8.2). The north-south valley between Plowden Station and the Clun is a continuation of the

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1. Op. cit., p. 234
2. Dwerryhouse, A.R. and A.A. Miller, 1930, p. 101-102

course of the East Onny, while the depression from Lydham Heath to Brockton bears a similar relationship to the upper part of the West Onny. During the glaciation a small glacier is believed to have flowed down the East Onny valley, at least as far as Plowden Station, and, as suggested by Harmer[1], the Severn ice blocked the mouth of the wide valley between Montgomery and Chirbury.

> "The effect of this would have been to divert much additional water into the Lydham Heath and Brockton valley, which may account[2] for its great breadth."

The further advance of the western ice closed the outlet of the Kemp valley to the Clun,

> "...thus forcing waters from both the West and East Onny to take a course along the present line of the Onny gorge from Plowden Station to Wistanstow and Craven Arms, which was, in our opinion, cut at this time."[3]

The retreat of the Clun glacier allowed the escape of the waters from the Kemp valley; but, as the Onny had already entrenched itself in its gorge, its waters were not restored to their old channel.

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1. Harmer, F.W., 1907, p. 481-483
2. Dwerryhouse, A.R. and A.A. Miller, 1930, p. 101
3. Ibid.

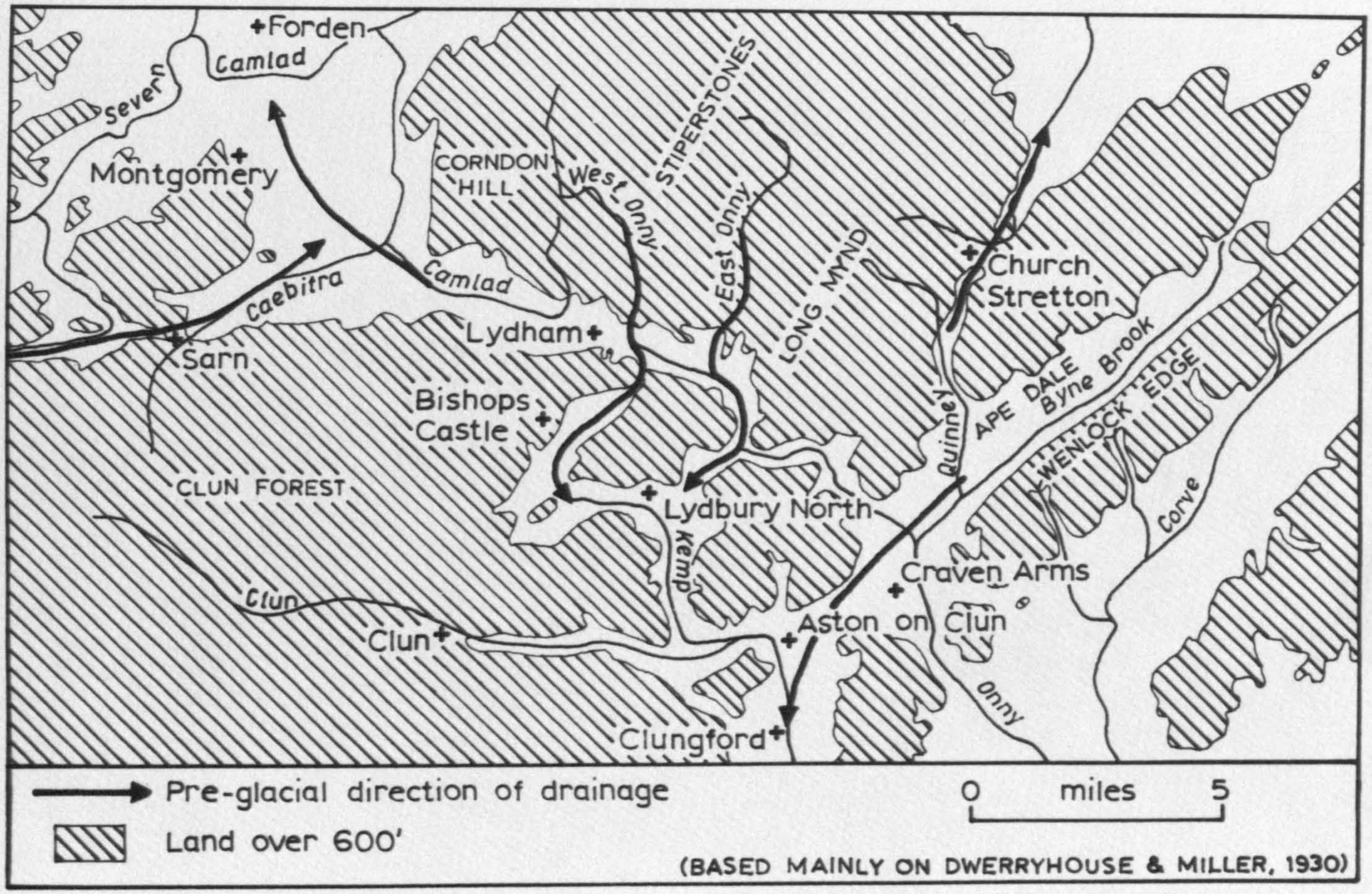


Fig. 8.2 Pre-glacial drainage, after Dwerryhouse and Miller

Harmer[1] contended that the Camlad originally flowed across the Wenlock Beds of the Montgomery-Chirbury gap, but was subsequently diverted to the Ordovician country to the east. Marrington dingle acted as an overflow channel to a small glacial lake to the south of Church Stoke, permanently diverting the drainage.

Dwerryhouse and Miller[2] also proposed that the pre-glacial course of the Mule was eastwards past Sarn into the wide, flat-bottomed Caebitra valley. They do not fully explain the diversion of the Mule, but state that, "The change in the direction ofthe stream is due to the blocking of its former course by morainic mounds."[3]

More recently, contrary views have been expresssed concerning the drainage diversions. Brown[4] attempts a reconstruction of the drainage pattern at the Low Peneplain stage (fig. 8.3), and is impressed by, the right-angled bend in the West Onny and the fact that the highest point in the Camlad-Onny valley is 600 ft at Lydham. This

> "...is probably to be explained in terms of an ancestral Onny which at the height of the col had its source north-west of the gap. It is of particular interest

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1. Op. cit.
2. Op. cit., p. 101
3. Ibid.
4. Brown, E.H., 1960, p. 132-133 amd 139-140

> that this proto-Onny is in direct line with the proto-Vyrnwy-Banw."[1]

The upper portions of the Vyrnwy and Banw flow from north-west to south-east, at right angles to the Caledenoid structural trend (fig. 8.1). They then turn sharply to the north-east and are carried to the Severn along the Vale of Meifod, below their junction.

> "There is evidence of a former more direct course of the joint upper Vyrnwy and Banw to the Severn, continuing their south-easterly trend via the lower Banw and the Castell Caereinion col at 586 ft. Capture by the present lower Vyrnwy in the Vale of Meifod was facilitated by erosion along the strike of the Llandovery rocks on the south flank of the Meifod anticline."[2]

The Low Peneplain extends through the Plowden gap and the proto-Onny is continued through this, across the Church Stretton Fault and the Wenlock and Aymestry Limestone scarps.

> "This seems to favour a proto-Vyrnwy-Onny which was not captured by the Severn until the rejuvination phase following the Low Peneplain stage. The Severn probably had by far the shorter distance to base-level and so the enrgy to effect the capture."[3]

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1. Op. cit., p. 132
2. Op cit., p. 131-132
3. Op. cit., p. 133

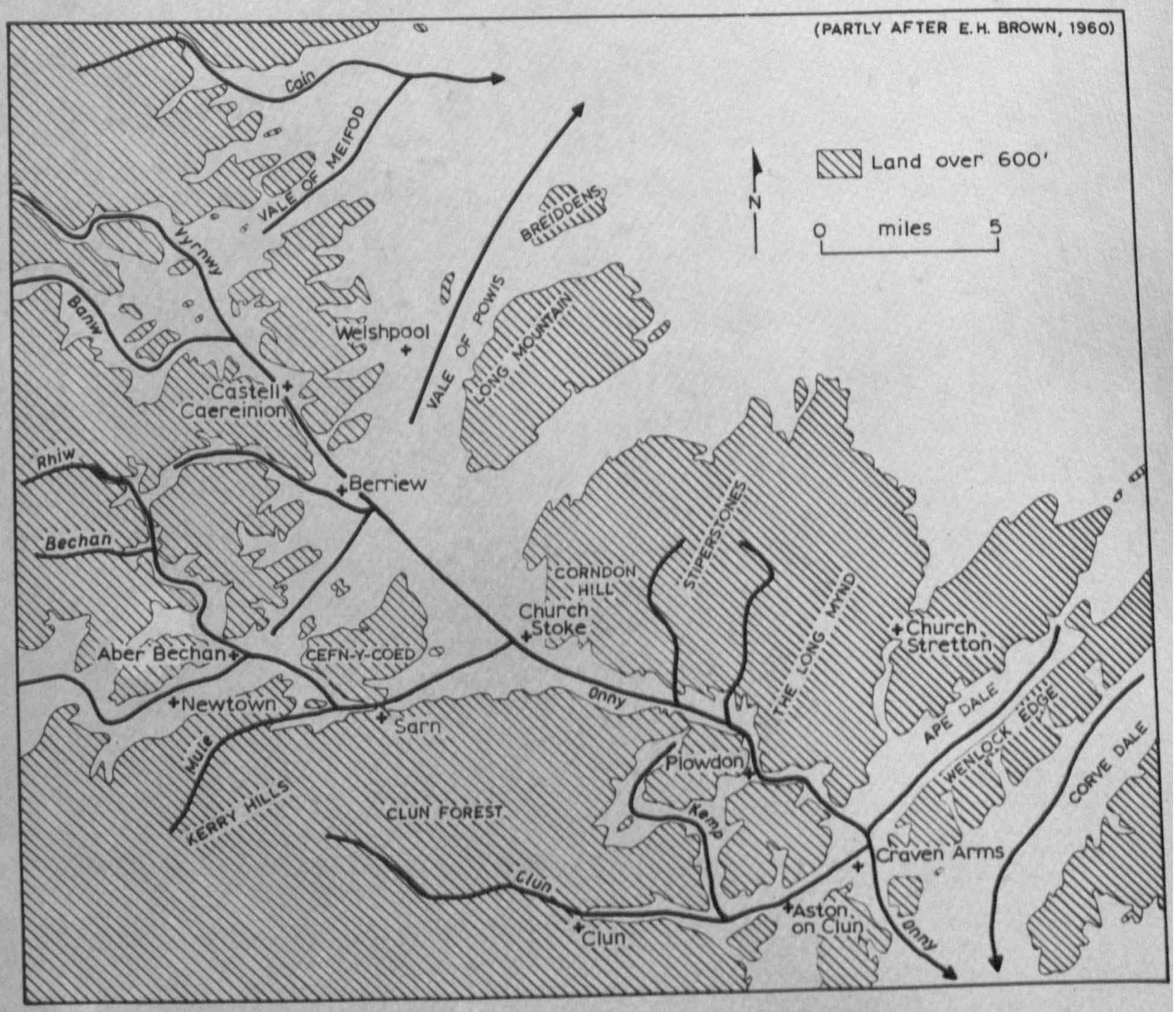


Fig 8.3 Drainage at the Low Peneplain stage

Brown doubts the reconstruction of the pre-glacial Onny drainage made by Dwerryhouse and Miller, which, "...cuts across the Camlad-Onny line which has tentatively been linked with the Vyrnwy on the Low Peneplain."[1] He also doubts that the Onny's present course around the southern end of the Long Mynd was initiated by the glacial diversion of a north-south stream, and that,

> "Judging by its form, the Lydham Heath-Brockton gap is more likely to be of a fluvio-glacial origin,"[2]

It is considered possible that, on the Low Peneplain, the Clun continued eastwards from Aston-on-Clun and passed through a gap, the floor of which is at 527 ft, to join the Onny at Craven Arms. The stretch between Aston and Craven Arms would have been a strike stream on the Wenlockian, a structural and geomorphological continuation of Ape Dale, east of the Onny.[3] On the other hand, the Soil Survey contends that the pre-glacial drainage of Ape Dale was carried in the opposite direction, to join the Clun at Clungunford.[4]

It would appear, therefore, that a number of conflicting views have been expressed regarding the diversions of

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1. Op. cit., p. 139
2. Ibid.
3. Op. cit., p. 140
4. Mackney, D. and C.P. Burnham, 1966, p.6

drainage in this area, and, apart from that of Brown, little detailed work has been carried out. Most authors explain the diversions in relation to former glacial lakes and their supposed overflow channels; most, especially that proposed by Cobbold which filled the entire Camlad-Onny valley, are highly improbable.

To elucidate the drainage diversions the pre-glacial drainage must first be reconstructed. This involves a knowledge of the depths of drift in the main valleys, a recognition of planation surfaces, with wind gaps and drainage alignments, and the relationships of the drainage to structure. It must also be made clear whether the diversions were effected by ice, by glacial meltwaters, or to what extent they pre-date the glacial period.

## ONNY DRAINAGE

The reconstruction of the pre-glacial drainage of the Onny and its tributaries is a difficult and complex task. There have evidently been several changes in the alignment of the streams since the Low Peneplain stage, not all, of which date from the glacial period. Both the Onny and the Quinney Brook are remarkable in that they flow through prominent gorges formed by glacial meltwaters,

cutting across the relief and geological structure. It is necessary to ascertain how far the erosion of the gorges was responsible for diverting the pre-glacial drainage, if at all. There are several drainage alignments and wind gaps at a similar elevation, suggesting various possible courses open to the Onny and its tributaries at the Low Peneplain stage (fig. 8.3). The East and West Onny could have continued southwards through the broad gaps on either side of Oakley Mynd, as suggested by Dwerryhouse and Miller. Alternatively, the proto-Vyrnwy-Onny could have continued eastwards through the Plowden gap, as suggested by Brown. The Quinney Brook may have flowed south-westwards past Cwm Head and Edgton to join the Kemp, cutting directly across the proto-Vyrnwy-Onny line. On the other hand, the Quinney may have continued southwards past Marshbrook, as at present, to join the proto-Onny.

Structural Relationships

Within the Onny basin, only the rocks to the east of Plowden have been mapped in any detail (fig. 1.7). The dominant structural feature of the area is the Church Stretton Fault complex, several branches of which pass through the broad gap between Marshbrook and Horderley, and past the Plowden gap at the southern end of the Long Mynd. Other

branches pass southwards, one along the Marshbrook gorge and another separating the Longmyndian of Hopesay hill from the Ludlow Series to the west. The western margin of the Long Mynd plateau is also dominated by a prominent fault scarp. In the south-east of the area the Ordovician and Silurian rocks strike north-east to south-west, parallel to the Church Stretton Fault. Alternating shales and limestones are expressed in the characteristic scarp and vale topography of the Wenlock Edge area.

The present drainage is strongly discordant to structure. The Onny, after passing through the Plowden gap, cuts directly across Longmyndian rocks at Horderley, and across Ordovician flags and shales; Below Craven Arms it traverses the prominent Aymestry and Wenlock limestone scarps. The Quinney Brook leaves the fault-guided Valw of Stretton and cuts across the strike of the Ordovician flags before entering Ape Dale, though its course is in part guided by a branch of the Church Stretton Fault.

Quinney Brook

Within the Vale of Stretton there are several erosion surfaces on the flanks of the Long Mynd, at heights varying between 800 and 700 ft. Above Church Stretton a spur is truncated at 800 ft., and Brockhurst hill has a level bedrock surface at 700 ft. To the north and west of

Little Stretton spurs adjoining Ashes Hollow and Callow Hollow are truncated at 750 ft. There are no such surfaces on the opposite side of the valley, but they may be traced southwards at Minton and on both sides of the Marshbrook gorge, which is incised into a broad, drift-covered platform at 700 ft.

The extrapolated long profiles of several of the deep valleys draining the eastern side of the Long Mynd grade to a 750 ft. level (fig. 8.4). The Upper part of Ashes Hollow is relatively broad and open, with narrow alluvial flats, and is separated from a lower mature stretch by a deeply-incised cross-section which is being eroded at the present day. A pronounced knick point in the long profile at Ashes Cottage, where the river is cutting into bedrock, corresponds with this deeply incised portion. The extrapolated profile of the upper part of the valley grades to the 750 ft surface on the spur north of Little Stretton. A similar relationship may be seen in Callow Hollow and Cardingmill valley.

It would appear, therefore, that there was a pre-glacial drainage level at about 750 ft in the Vale of Stretton, and that a rejuvination of the Long Mynd valleys occurred with the deepening of the main valley.

In direct alignment with the Vale of Stretton is the broad Cwm Head gap, lying between Whittingslow Ridge and the

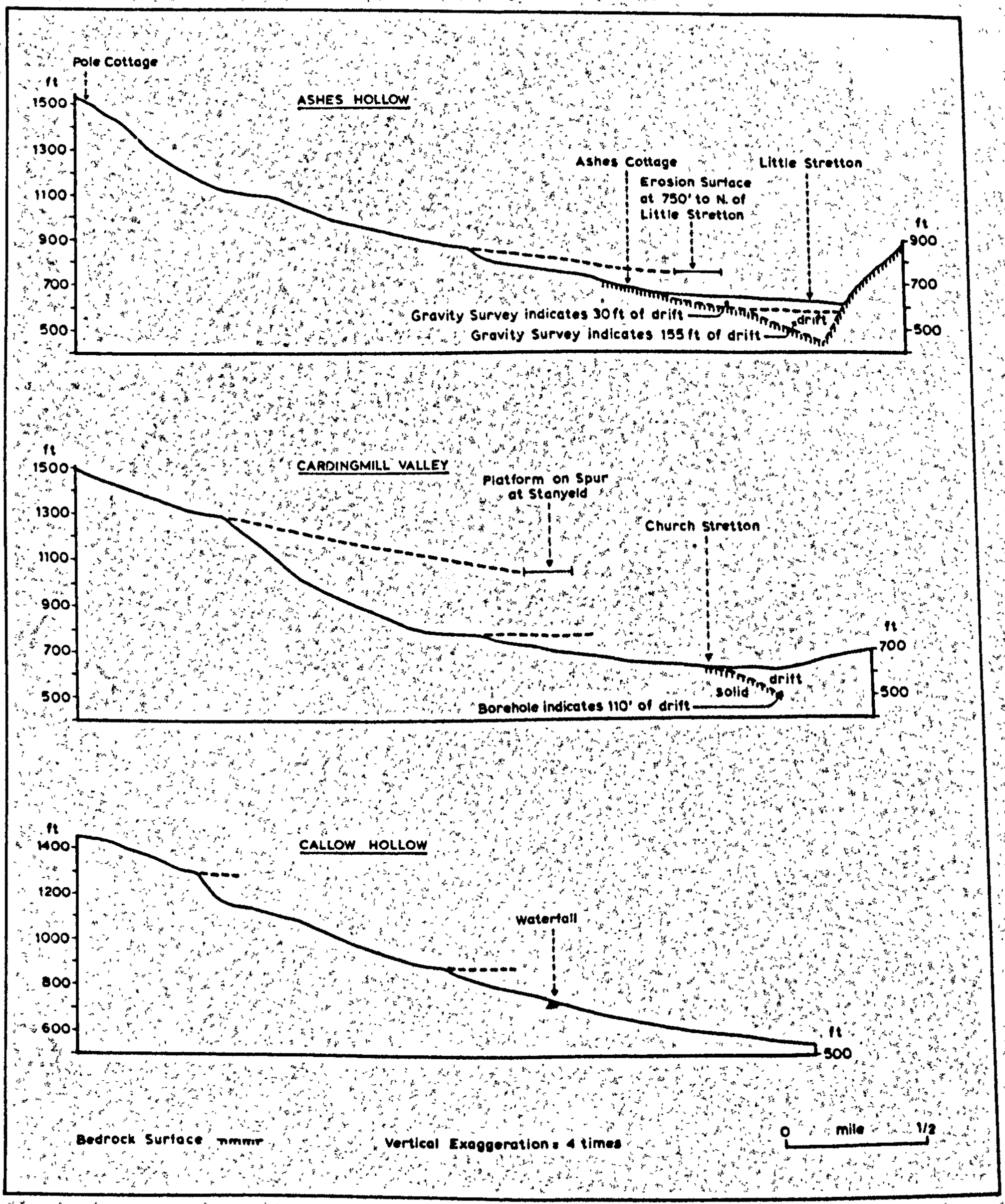



Fig. 8.4 Long profiles of rivers in the Vale of Stretton

Long Mynd. The gap is at present drained by two streams flowing in opposite directions, separated by a dry col at Cwm Head at a height of 762 ft. This line is continued by a broad col at Edgton on the opposite side of the Onny valley, at 768 ft.

It is probable that a proto-Quinney flowed south-eastwards, past Minton, through the gaps at Cwm Head and Edgton and became tributary to the Kemp. This would be a subsequent stream, flowing parallel to the regional strike. It would also flow parallel to several other major valleys in the area, including Ape Dale, Hope Dale and Corve Dale, all of which are subsequent. The prominent gap near Sibdon Carwood and Long Meadowend, the floor of which is at 500 ft, possibly carried the drainage of Ape Dale into the Clun near Clungunford (fig. 8.5).

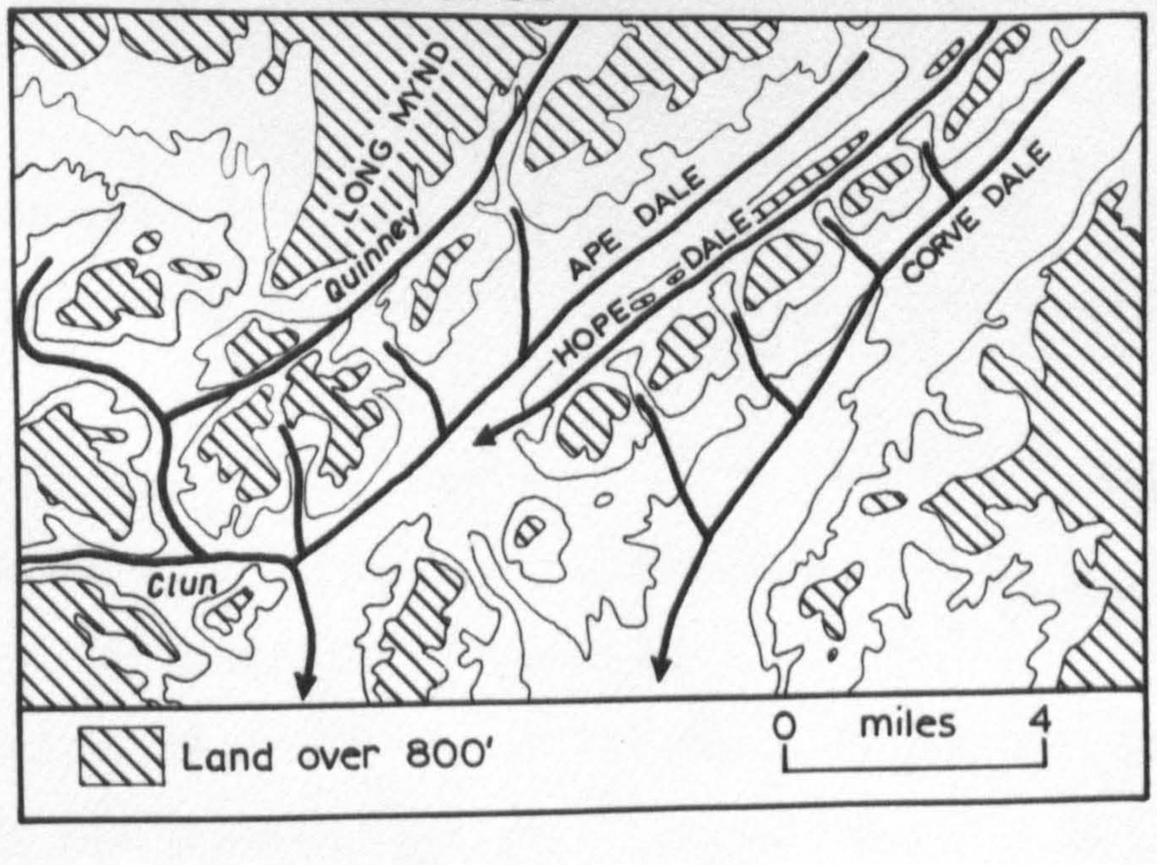
The platform on either side of the Marshbrook gorge is at 700 ft, which is lower than most surfaces in the Vale of Stretton, and the Cwm Head and Edgton gaps. It is likely, therefore, that drainage past Marshbrook was initiated at a slightly later stage than that through the Cwm Head gap. There may have formerly been a stream flowing down the southern flank of the Whittingslow ridge, down the dip of the strata. This would be able to extend itself headwards along the line of weakness associated with the Church Stretton Fault; it would eventually breach the

ridge and capture the proto-Quinney, leaving the Cwm Head gap dry. At Edgton there is a deep valley running southwards past Hopesay into the Clun. This was evidently cut by a vigorous stream flowing down the dip of the Ludlow shales and partly along a fault line. The stream would extend itself headwards and capture the lower part of the proto-Quinney, leaving the Edgton gap dry.

A similar state of affairs prevailed in the scarp and vale topography of Wenlock Edge. Hope Dale is characterised by a series of dry gaps which fall in height from about 750 ft in the north-east to 700 ft in the south-wast. These were cut by a river which flowed parallel to the proto-Quinney at the 750 ft stage, eroding a valley in the Lower Ludlow shales. Several small streams flowing down the dip of the Upper Ludlow Shales eventually breached the Aymestry Limestone scarp bordering the valley, and disrupted the drainage by a series of captures (fig. 8.5).

Thus at the 750 ft stage there were several major rivers, including the proto-Quinney, which flowed from north-east to south-west. Subsequently a series of river captures, operating on a regional scale, disrupted the drainage and diverted it through major scarps and ridges. The captures were effected by relatively vigorous streams flowing down the dip of the strata, in some cases exploiting

DRAINAGE AT 750' STAGE




DRAINAGE AT 700' STAGE (Pre-glacial drainage)

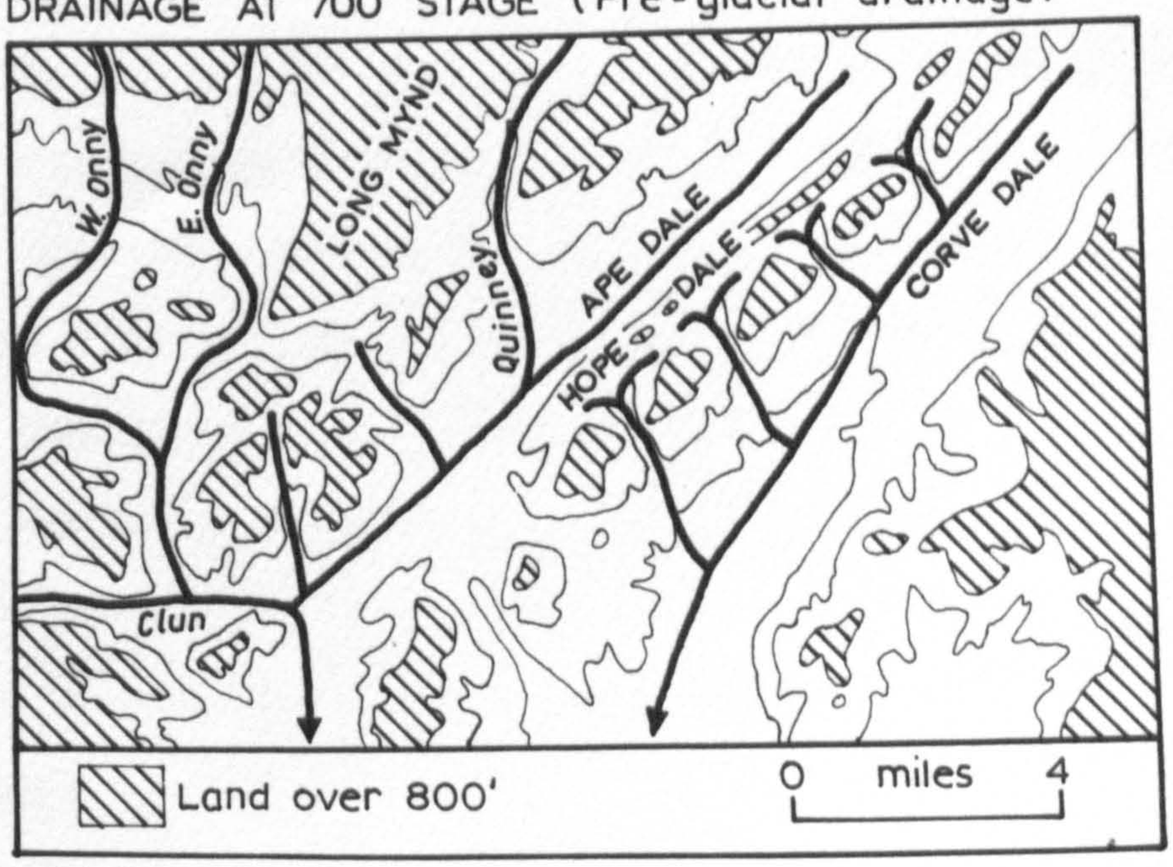



Fig. 8.5 Development of pre-glacial drainage

lines of weakness associated with faults. Subsequent streams were replaced by discordant drainage below the 750 ft stage.

There is evidence of drainage at a still lower elevation in the Vale of Stretton in pre-glacial times. Borehole records and temporary excavations indicate a sub-drift rock platform at 600 ft to the north of Church Stretton, though the sub-drift surface descends rapidly to 500 ft. on the eastern side of the valley, the overdeepening probably being due to glacial erosion. The extrapolated long profile of Cardingmill valley (fig. 8.4) grades to a 600 ft level. Similar relationships prevail opposite Ashes Hollow. At Ashes Cottage the stream flows for a short distance over a bedrock floor,and there is about 30 ft of drift at the mouth of the valley. Beyond this the sub-drift surface deepens rapidly to about 450 ft on the eastern side of the Vale of Stretton. Again, this overdeepening is probably the result of glacial erosion. The extrapolated long profile of the bedrock floor of Ashes Hollow grades to a level of about 575 ft.

Remnants of a pre-glacial valley may also be found in the Marshbrook gorge. There is a marked steepening of the cross profile on the eastern side of the gorge, below 600 ft (fig. 8.6), indicating the incision of a meltwater

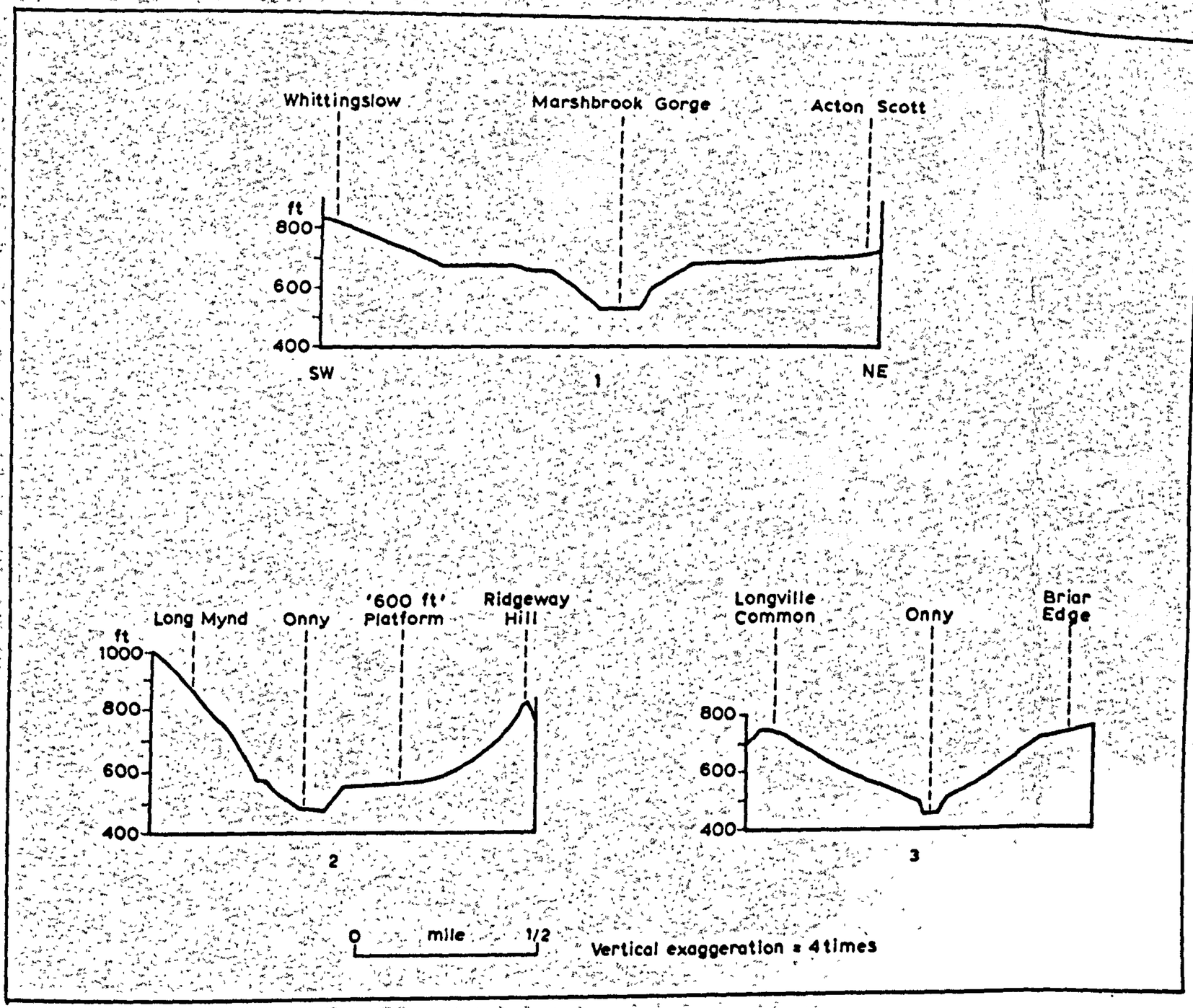



Fig. 8.6 Cross profiles of the Plowden and Marshbrook gorges. Position of profiles is shown on Fig. 8.7

channel into a former valley floor.

Since there has been considerable glacial overdeepening in the Vale of Stretton and the bedrock profile is so irregular, it is difficult to accurately re-construct the pre-glacial valley. However, the evidence of the sub-drift platforms, extrapolated bedrock profiles of rivers, and the 'valley-in-valley' form of Marshbrook gorge, point to a former drainage level at about 600 ft. Since the Quinney Brook had already been diverted past Marshbrook in pre-glacial times, the erosion of the gorge by glacial meltwaters did not effect any re-alignment of the drainage. It merely re-emphasised a pre-existing drainage line.

Upper Onny Basin

The reconstructed proto-Quinney at the 750 ft stage cuts directly across the line of the proto-Vyrnwy-Onny proposed by Brown. On the other hand, it is possible that the East and West Onnys continued southwards through the gaps on either side of Oakley Mynd and entered the Clun near Clunbury. Both gaps are covered with glacial drift, and their floors are at heights of 525 ft and 560 ft respectively. Brown partly based his reconstruction on the height of the watershed at Lydham, which is at 600 ft. But, the Onny valley at this point is thickly covered with drift, and a kame moraine comprises the watershed. In order to

re-construct the pre-glacial drainage it is necessary to determine the height of the sub-drift surface within the valleys.

At More, opposite the 'elbow' in the West Onny, 15 ft. of gravels overlies boulder clay to a depth of 35 ft. Unfortunately the absolute depth of drift is not known at this point. At Oakley Farm, on the eastern side of the gap opposite the West Onny, a bore proved 101 ft of glacial drift, and bedrock must lie below 469 ft. At Brockton, at the southern end of the gap, 24 ft of drift overlies bedrock, the surface of which is at 476 ft. While at Lydbury North, on the northern side of the Kemp valley, 12 ft of drift rests on bedrock, the surface of which is at 513 ft.

The sub-drift surface of the Lydham Heath-Brockton gap is therefore below 470 ft. The bore at Brockton was sunk at the edge of the valley and does not, therefore, necessarily indicate the deepest portion, whereas that at Oakley Farm was close to the valley centre.

The depth of drift is also known at the southern end of the East Onny valley. At Hardwick 62 ft of drift rests on bedrock, the surface of which is at 500 ft. On the opposite side of the river, several bores at The Home indicate a maximum of 25 ft of drift, and the bedrock surface is at 500 ft. At Eaton, immediately below the confluence of the East and West Onnys, 25 ft. of drift rest on a bedrock

surface at 525 ft.

There are no borehole records for the gap on the eastern side of Oakley Mynd. The valley floor is covered by alluvium amd drift, and the northern end is truncated by the Onny gorge, which exposes bedrock on its side. The height of the bedrock beneath the valley floor is therefore at 525 ft. A prominent spur extends south down the centre of the gap; it has a flat bedrock surface at 600 ft.

Between Plowden and Horderley the Onny gorge is aligned roughly west-east, coincident with the strike of the Hughley shales and Pentamerus beds which rest unconformably on the Pre-Cambrian rocks of the Long Mynd. The dominant feature of this section of the gorge is a broad rock platform on the southern side at 600 ft, which slopes towards the centre of the valley at angles varying between $1^{o}$ and $4^{o}$. Fragments of the platform are also preserved on the Long Mynd side of the gorge. The platform is backed on the south by a steep scarp over 200 ft high, which near Horderley appears to truncate Ridgeway hill. The Onny is incised into a broad gorge some 25 ft deep below the platform, which has been cut by glacial meltwaters.

Below Horderley the Onny valley is markedly different. It is narrower and is discordant to structure, cutting across Pre-Cambrian rocks, Ordovician flags and shales, and the

Church Stretton Fault. Both the 600 ft platform and the steep scarp edge are absent. Cross profiles indicate a 'valley-in-valley' form, a meltwater channel having been incised into the floor of a pre-existing valley at about 500 ft. (fig. 8.6).

From the height of the sub-drift surface it seems probable that the West Onny continued southwards through the Lydham Heath-Brockton gap in pre-glacial times, with the valley floor lying below 470 ft. The sub-drift surface of the East Onny at its junction with the West Onny is at 525 ft, which is the same height as the bedrock valley floor in the gap to the east of Oakley Mynd; whereas the rock platform in the Plowden gap is at 600 ft. It appears, therefore, that the pre-glacial East Onny flowed south through the gap east of Oakley Mynd, rather than eastwards past Plowden.

It is possible that the proto-Onny passed through the Plowden gap at the 600 ft. stage. But, also, the Onny could have comtinued southwards and cut the 600 ft. surface on the spur in the gap east of Oakley Mynd.

It has been suggested that the proto-Quinney, which cuts across the line of Brown's proto-Vyrnwy-Onny, was disrupted by captures at Edgton and Marshbrook. It is also possible that a stream flowed southwards down the dip of the strata in the valley now occupied by the Onny gorge

below Horderley. This stream would also be operative in the capture of the Quinney Brook below the 750 ft stage.

It seems, therfore, that there was no pre-glacial drainage through the gap between Plowden and Horderley. The gap may, instead, have been cut by glacial erosion, a transfluence col formed by ice breaching a col at 900 ft at the southern end of the Long Mynd. Ice passing eastwards down the Camlad-Onny valley was subjected to pressure from the south by the Clun glacier (fig.8.7 ). This would effectively prevent it from passing southwards through the gap to the east of Oakeley Mynd, but would force it northwards or north-eastwards against the Long Mynd. A considerable volume of ice moved northwards up the East Onny valley, and other ice passing north-eastwards round the southern end of the Long Mynd would be effective in carving the Plowden gap. The latter closely resembles a glacial trough, though the more resistant Langmyndian rocks would mitigate against the cutting of a steep wall on the northern side. The ice, passing round the southern end of the Long Mynd would be following a line of least resistance, and would effectively erode a trough along the strike of the relatively soft Hughley shales and Pentamerus beds. Beyond Horderley, the ice continued through the Cwm Head gap towards Little Stretton.

With the onset of deglaciation the ice gradually

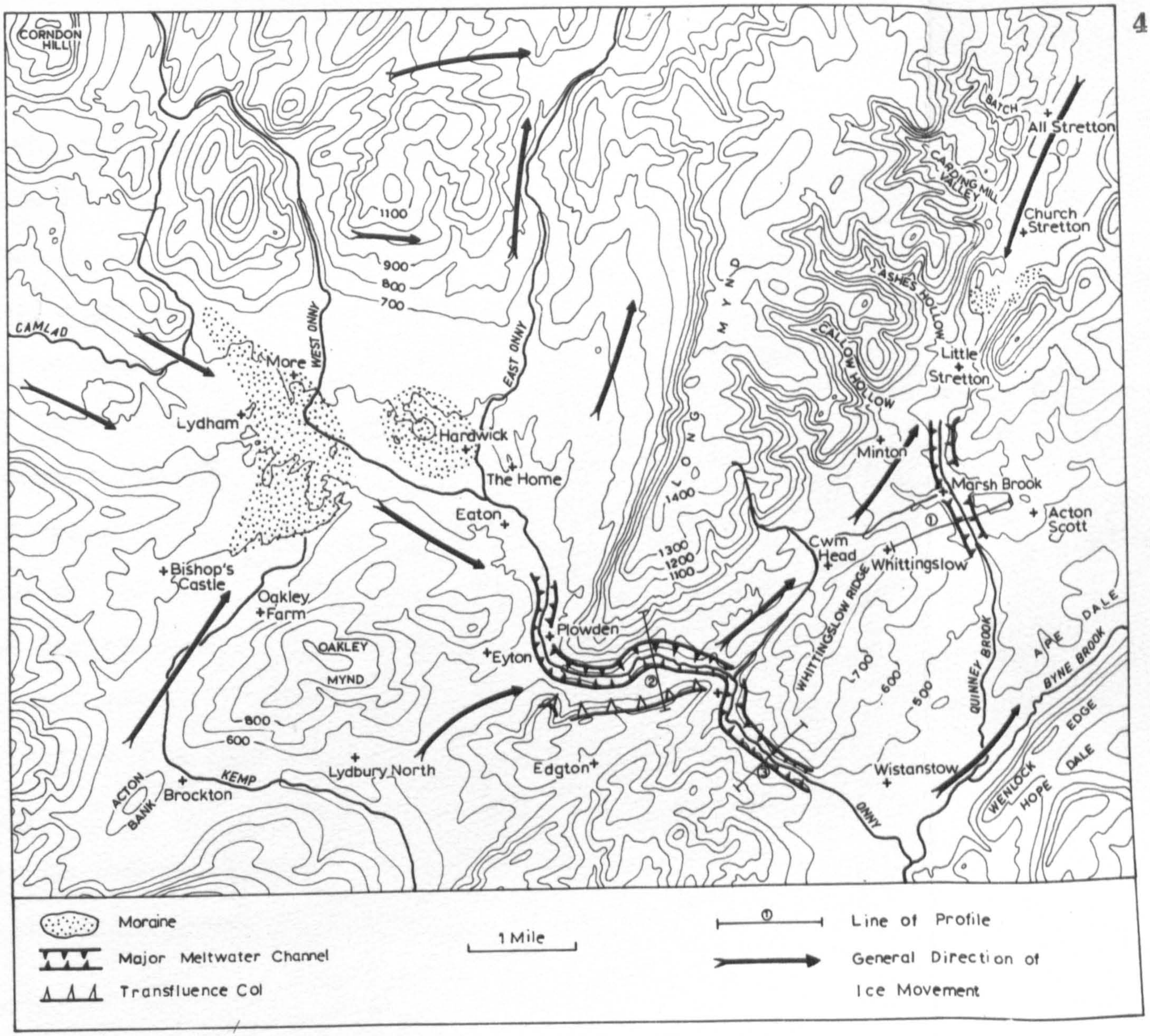



Fig. 8.7 Ice movements at the southern end of the Long Mynd

downwasted, and subglacial channels were cut at Marshbrook, within the Cwm Head gap, and in the Onny gap between Plowden and Wistanstow. The latter was incised below the floor of the pre-glacial valley to the east of Oakley Mynd, and after the disappearance of the ice the drainage was permanently diverted through the channel. Thus, although the channel at Marshbrook re-emphasised a pre-glacial drainage line, that in the Plowden gap was respomsible for a major re-alignment of the drainage.

The deposition of the kame moraine at Lydham, and the considerable volume of drift in the Lydham Heath-Brockton gap, effectively blocked the pre-glacial course of the West Onny. With the retreat of the ice from Lydham the West Onny was no longer able to flow through the gap, and turned abruptly eastwards at More.

## CAMLAD DRAINAGE

An interesting example of river capture occurs in the headwaters of the Camlad and West Onny (fig. 8.9). From its source in Black Marsh the Onny flows south over a broad, drift-covered valley. At Runnis Bridge the river makes a sharp turn to the north-east and joins a stream flowing southwards from the Shelve area. Below the point of confluence the river continues south through a deep,

steep-sided valley east of Heath Mynd towards Lydham. South of Runnis Bridge the upper part of the West Onny is directly continued by a large valley between Cefn Gunthley and Bank. A col at Pulthely, at 850 ft, in the centre of this valley, separates a northward-flowing tributary of the West Onny from the headwaters of the Camlad. At the col is an accumulation of drift, the thickness of which is not known.

The southward-flowing sections of the streams are parallel to the direction of the geological strike. The upper part of the West Onny flows along Mytton flags and shales before cutting across a volcanic ridge at Runnisgate.

It is probable that the Camlad originally continued southwards past Runnis Bridge, and that the West Onny had its source in the Shelve area. The north-eastward flowing section of the West Onny was probably a small tributary which captured the headwaters of the Camlad. This may have been a normal river capture, effected by a stream extending itself headwards along the Ordovician shales. Glaciation may also have played a large part in the capture. There was a general east and north-east movement of ice from the Corndon hill area towards the Long Mynd, and from the Cefn area towards the Shelve. The basal layers of the ice would tend to follow the valleys, and it is likely that the valley north-east of Runnis Bridge was deepened

and a pre-glacial watershed breached by ice moving towards Shelve from Cefn. The deposition of drift at Puhthely created a watershed in the original Camlad valley. With the retreat of the ice the headwaters of the Camlad would be permanently captured by the West Onny.

Below Lydham the Camlad pursues an interesting circuitous route to the Severn valley. From the low watershed formed by the moraine at Lydham it flows north-westwards towards Church Stoke, where it joins the eastward-flowing Caebitra (fig. 8.8). The combined waters pass northwards through Marrington dingle into the Marton valley. Morainic drift blocks the latter at Hockleton, and the Camlad turns sharply westwards and enters the Severn near Forden.

Brown proposed that at the Low Peneplain stage a proto-Vyrnwy-Onny passed across the Montgomery-Chirbury gap towards Plowden. But, since the West Onny flowed south through the Lydham Heath-Brockton gap at this stage the Vyrnwy-Onny must have joined the West Onny near Lydham. Also, as will be shown below, a proto-Mule-Caebitra flowed eastwards past Sarn and met the Vyrnwy-Onny at Church Stoke . Immediately prior to glaciation, however, the pattern was somewhat different. The subsequent river Severn captured the Vyrnwy-Omny near Berriew during the rejuvination

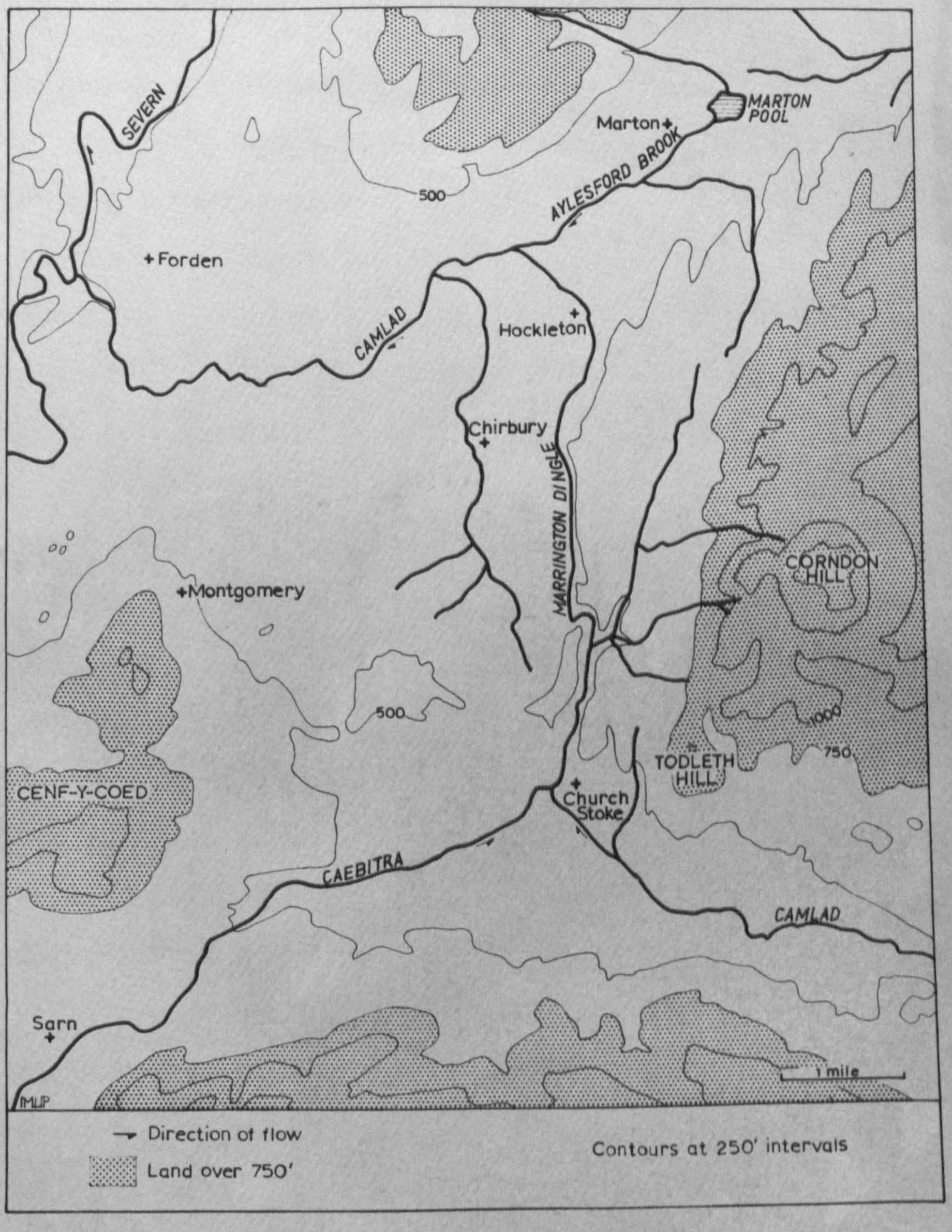



Fig. 8.8 Caebitra - Camlad drainage pattern

phase following the Low Peneplain stage[1], and it is likely that drainage directly across the Montgomery-Chirbury gap was terminated (fig. 8.9).

Apart from Marrington dingle, the gap forms a watershed between the Marton and Caebitra-Camlad valleys and probably functioned as such in pre-glacial times. At Moat and Timberth between 7 and 8 ft of till overlie bedrock, the surface of which is at 442 ft. At Rhiston there is only 2 ft of soil and the rock surface is at 500 ft. This is the approximate height of the pre-glacial watershed.

Certain elements of a pre-glacial drainage pattern are evident within the gap. West of Old Church Stoke the predominant direction of the drainage is towards the south, and the Coed Brook enters the northward-draining Marrington dingle at Calcot. To the south the dingle is apparently incised into the floor of a southward-draining pre-glacial valley. From Church Stoke to Calcot the ground level gradually rises, though the floor of the dingle progressively deepens. A series of profiles across the dingle (fig. 8.10) suggest that the pre-glacial valley floor was at about 500 ft near Calcot (profile I-J and E-F), and fell to about 440 ft north of Church Stoke (profile A-B). Here the entrance to the dingle is incised into a broad, drift-covered platform, the surface of which is at about 425 ft. This platform may be traced along the northern side of the

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1. Brown, E.H.: 1960, p. 133

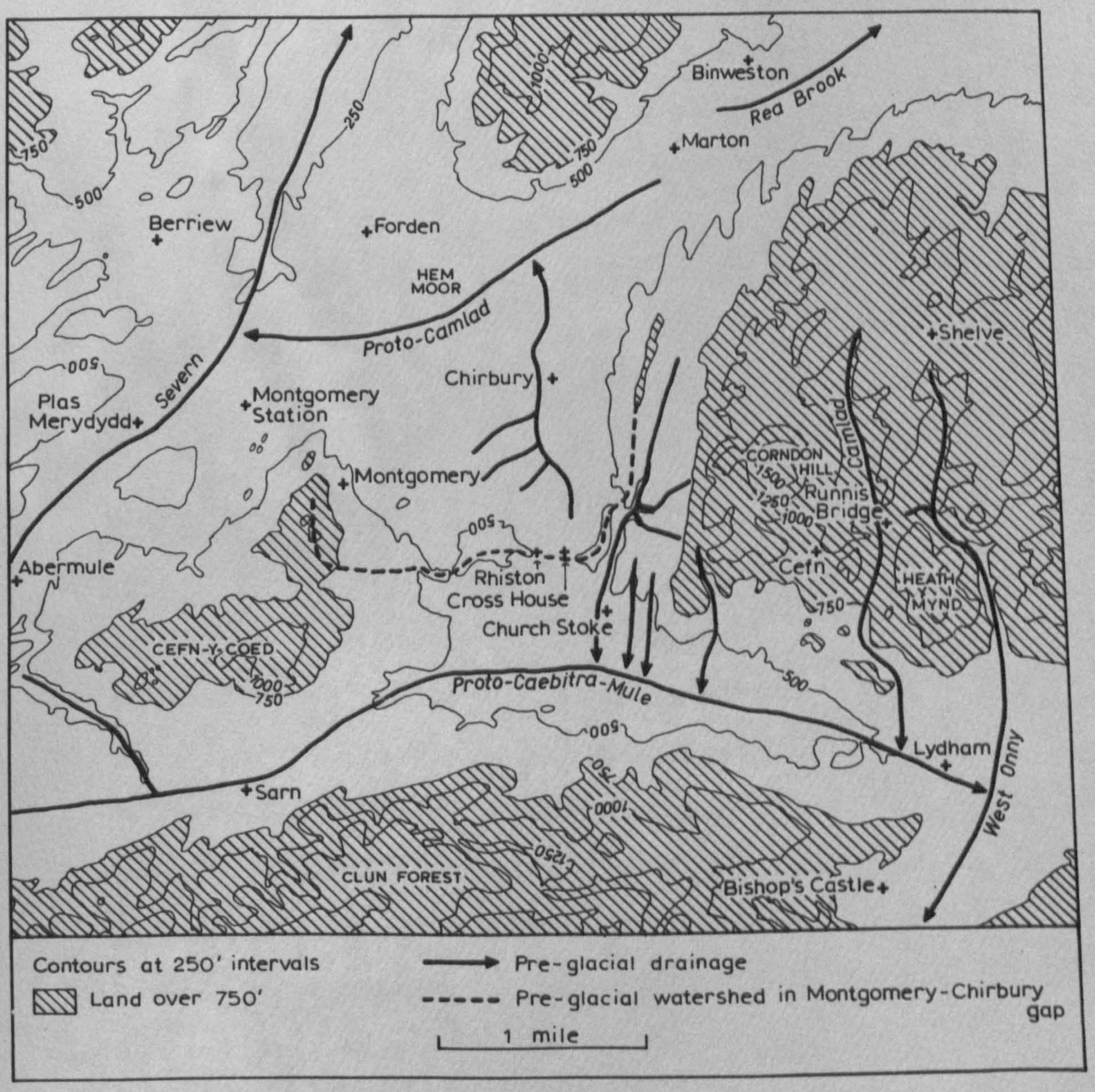


Fig. 8.9 Pre-glacial drainage pattern

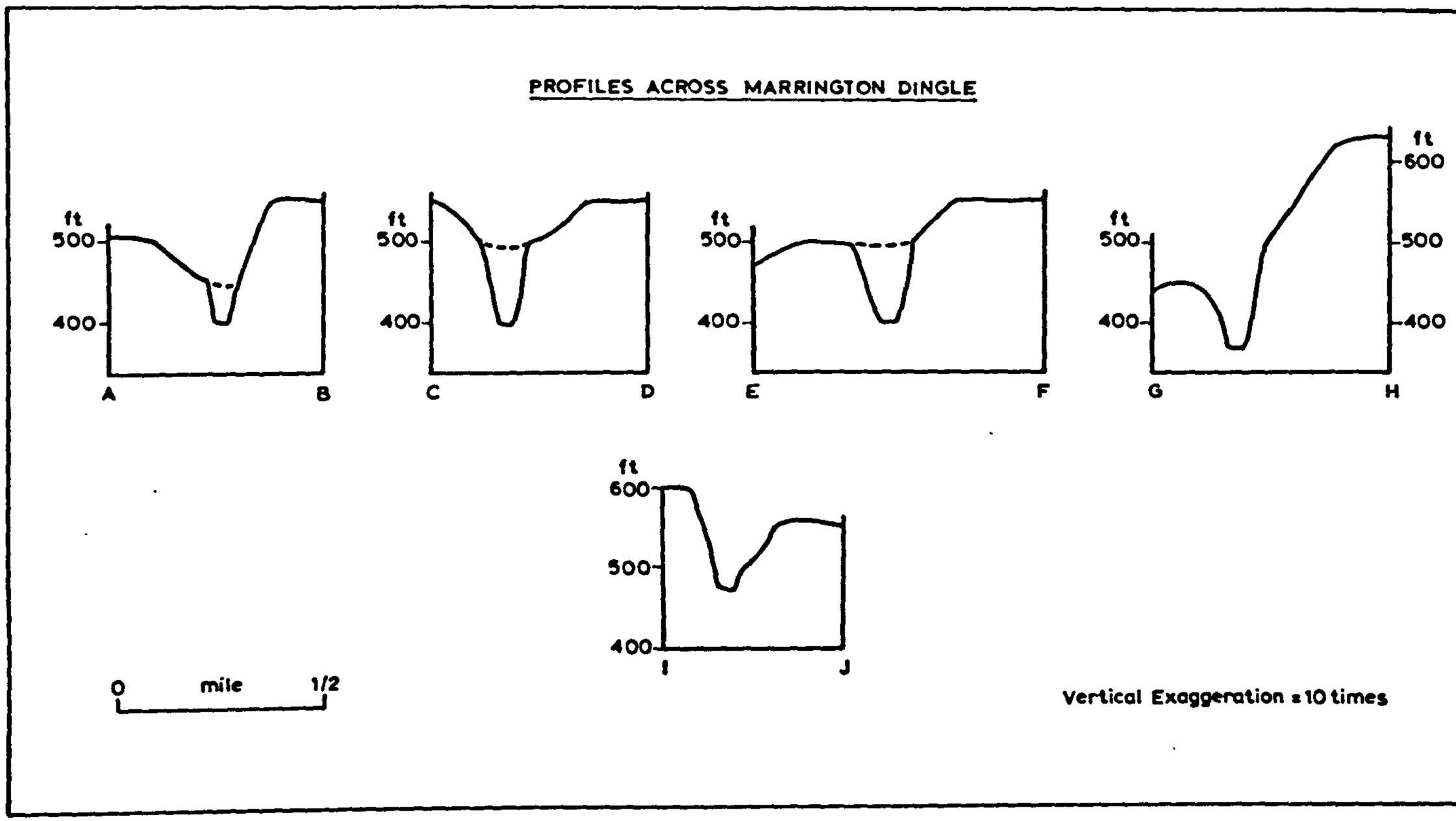


Fig. 8.10 Cross profiles across Marrington Dingle

Camlad valley, and is possibly part of the pre-glacial valley floor.

At Calcot, Marrington dingle is incised across a ridge of Hagley volcanics which forms part of the pre-glacial watershed in the gap. Beyond the ridge the dingle is a asymmetrical, the eastern side being much higher than the western, and it does not occupy a pre-glacial valley. The channel has been incised into the edge of the higher Ordovician country on the eastern side of the gap.

Several valleys trend northeastwards from Lymore Park towards Chirbury. They are tributary to a broad northward-draining valley which has its source to the west of Marrington. Although possibly modified by the passage of ice over them, the valleys are essentially part of a pre-glacial drainage system which carried waters into the Marton valley. At Chirbury 38 ft of drift overlie a bedrock surface at 312 ft. It would seem, therefore, that the pre-glacial valley graded to the Marton valley at a height of about 300 ft.

In pre-glacial times the Montgomery-Chirbury gap was drained by streams that flowed both north and south into the Marton and Caebitra-Camlad valleys. The pre-glacial watershed ran along the volcanic ridge now breached by Marrington dingle and along the southern end of the gap at a height of about 500 ft.

Major changes of drainage were effected during deglaciation. Meltwaters flowing along the edge of the ice against the Ordovician hills on the eastern side of the Montgomery-Chirbury gap eroded Marrington dingle. Meltwaters were directed northwards across the gap, flowing across the pre-glacial watershed. With the retreat of the ice the Plowdeh gorge was replaced by Marrington dingle as the major outlet for meltwaters in the Caebitra and Camlad valleys. At Hockleton Bridge the meltwaters originally escaped eastwards along the Marton valley. However, the ice evidently paused in its retreat at Hockleton and deposited kames and fluvio-glacial material across the valley. With the further retreat of the ice the moraine would form an effective barrier to the meltwaters. A small lake may have been temporarily impounded between the ice front and the moraine, but when the ice feached the Severn valley the meltwaters would be able to escape northwards to the Shropshire Plain.

After the disappearance of the ice the Lydham moraine remained as a barrier to the eastward flow of the Camlad, and, together with the Caebitra, it was permanently diverted through Marrington dingle. The Hockleton moraine was also a barrier and the Camlad now turns west towards the Severn, The diversion of the Camlad is thus attributed to the blocking of pre-glacial valleys by drift barriers and the

incision of a major meltwater channel across the pre-glacial watershed.

## MULE DRAINAGE

Between Kerry and the moraine at Sarn the Mule flows in a broad valley aligned roughly east-west (fig.3.4 ). The valley is continued eastwards, past Sarn, by the broad Caebitra valley. The moraine forms a watershed at 600 ft between the two. Both valleys are underlain by relatively soft Lower Ludlow shales, and are bounded to the north and south by harder grits and sandstones forming the Cefn-y-Coed hills and Kerry hill. The two valleys thus comprise a continuous feature, both structurally and geomorphologically.

The remnants of several planation surfaces occur on the northern spurs of the Kerry hills, each of which may be traced continuously from the Mule into the Caebitra valley, and correspond to Brown's Middle and Low Peneplains. Since the erosion surfaces of Wales were created by fluvial erasion[1] there was probably an eastward-flowing river to the north of the Kerry hills, at least until the Low Peneplain stage. The proto-Mule-Caebitra post-dates

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1. Brown, E.H., 1960, p. 123-157

a southward-draining proto-Carno-Clun, which passed over the Kerry hills through a col at 1,450 ft to the west of Kerry Pole[1]. The river flowed discordantly to the predominent Caladenoid trend and was probably captured by the proto-Mule-Caebitra which flowed along the strike of the rocks.

Below the prominent 'elbow',at Sarn the Mule flows in a wide north-west to south-east valley before entering the Mule gorge at Llanmerewig. The valley is in direct alignment with the Bechan valley on the opposite side of the Severn, which is separated by a dry gap at 584 ft between Tregynan and New Mills, from the upper part of the Rhiw (fig. 8.1) It seems likely that the Rhiw flowed south through this gap at the Low Peneplain stage, and the proto-Rhiw-Bechan may have continued across the Severn into the Mule, over a col at 500 ft at Llanmerewig. The proto-Rhiw-Bechan was thus tributary to the proto-Mule-Caebitra on the Low Peneplain. The upper Rhiw was subsequently captured by the present lower Rhiw at New Mills, and the Severn diverted the Bechan Brook at Aber Bechan. The section of the Mule above the gorge remained as a beheaded tributary to the proto-Mule-Caebitra.

It must now be established whether or not the Mule

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1. Brown, E.H., 1960, p. 144-145

continued to flow eastwards past Sarn immediately prior to glaciation. The Mule valley is covered with thick deposits of drift, a bore having proved 100 ft at Kerry and 104 ft at Pen-y-gelli. The bedrock surface is at or below 580 ft and 460 ft respectively. Unfortunately, it has not been possible to determine the depth of the drift in the gap at Sarn. At Llwyncowrid there is over 50 ft of drift and bedrock is below 550 ft. To the east of Sarn there is also 50 ft of drift and bedrock must be at or below 550 ft.

The long profile of the Mule also gives some idea of the depth of drift (fig. 8.11). The steep upper part of the river flows mainly over a bedrock floor, though the morainic accumulation at Black Gate is prominent. Where the river opens out onto the wide valley at Kerry there is a distinct flattening of the profile, though there is a knick point at Glan Mule, upstream of the'elbow', and a less-marked steepening at the elbow itself. Below this, the river continues at a steeper gradient to the Severn. The approximate bedrock profile of the Mule valley has also been constructed. The gradient is roughly the same as the present river, and if it is extrapolated through the Sarn gap over 200 ft of drift should be present.

The long profile of the upper Caebitra has also been drawn (fig. 8.12). The valley is narrow and steep-sided, and the river flows with a steep gradient over bedrock for

Fig. 2.11

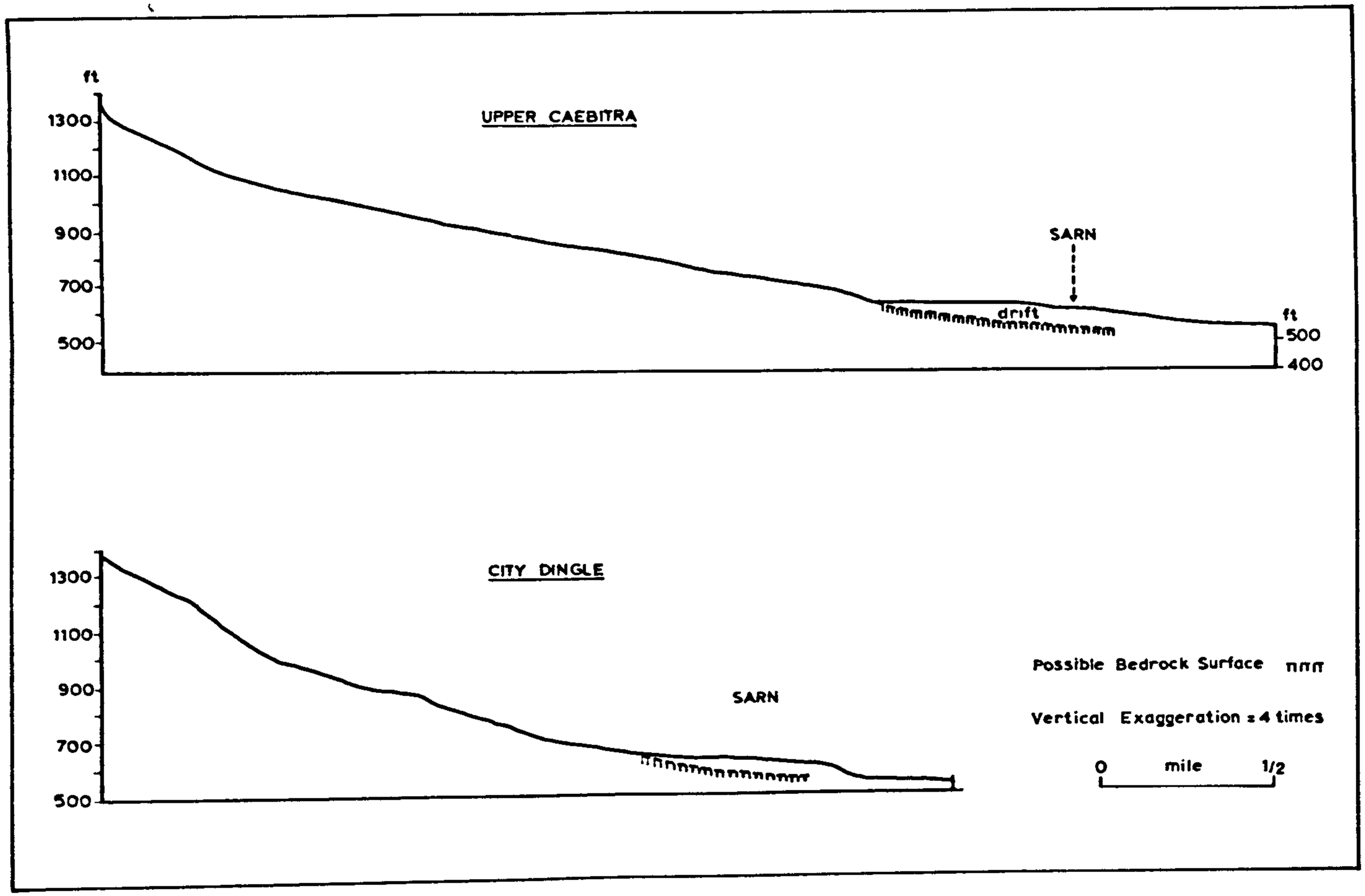


Fig. 8.12 Long profiles of the Caebitra and City dingle

much of its length. There is a marked flattening of the profile at about 600 ft where the river opens out onto the broad lower Caebitra valley. If the steep upper portion of the profile is extrapolated it indicates only about 80-100 ft of drift in the Sarn gap.

The long profile of City Dingle, a tributary of the Caebitra, shows a similar relationship. There is a marked flattening of the profile below 700 ft. and the extrapolatiom of the steep upper portion indicates about 100 ft of drift in the gap.

Only 50 ft of drift have been proved by bores at Sarn and Llwyncowrid, though the extrapolated profiles of the Caebitra and City Dingle give an estimation of about 100 ft. This is considerably less than the extrapolated profile of the Mule, the bedrock floor of which is about 100 ft below that in the Sarn gap.

If the Mule continued past Sarn prior to glaciation, the deeper portion of the valley above Sarn can only be explained in terms of glacial overdeepening. A considerable volume of ice filled the Mule-Caebitra valley, and must have been about 1000 ft thick at the glacial maximum. The ice streamlined numerous bedrock surfaces and formed two drumlin fields as it advanced, and it is not improbable that a considerable amount of glacial overdeepening took place, Both glacial erosion and deposition have considerably modified the long profile of the Mule.

Below the 'elbow' there is a marked contrast between the gorge section and the rest of the Mule valley. The gorge is a narrow, steep-sided feature, and cuts directly across the relief. It has been incised into the floor of a col at 500 ft west of Llanmerewig, and was formed by the superimposition of englacial meltwaters onto the underlying relief. The gorge was probably cut when the ice front stood at the Sarn moraine, and carried meltwaters from the upper Mule basin into the Severn valley.

When the ice retreated the Sarn moraine obstructed the eastward course of the Mule, and the river was permanently diverted through the gorge, leaving the moraine as a watershed. The new course of the Mule has the added advantage that it is much shorter to base level.

## SEVERN DRAINAGE

The Severn is a subsequent river, its valley has a marked Caledenoid trend, and it is possible that it is developed along a series of anticlinal axes. The Low Peneplain is well-developed throughout the Severn valley and its tributaries. Brown deduced that the Severn was fragmented at the Low Peneplain stage, and was crossed in several places by streams flowing transverse to the structure (fig. 8.3). The present Severn owes its existence to a series

of captures effected during a rejuvination phase following the Low Peneplain stage. The river was able to extend itself headwards along the geological strike and capture the transverse drainage.

Below Newtown both sides of the valley are characterised by a series of low platforms at various elevations below 600 ft. The lowest stands 25-30 ft above the valley floor, at 325 ft, and may be traced intermittently to beyond Montgomery Station. Remnants of the platform occur on both sides of the narrow ridge which extends along the centre of the valley from Plas Meredydd. The river cuts through it in a narrow gorge at Bryn-llyn.

A broad undulating platform at a similar height borders the Severn and Marton valleys west and south of Forden. The platform is less evident on the eastern side of the Severn beyond Forden, but it occurs intermittently on the opposite side of the valley. Remnants also extend northeastwards up the Marton valley, and the mouth of Marrington Dingle is incised across it. The platform rises gradually up the Marton valley, to about 350 ft. beyond the watershed.

The platform probably comprises the remnants of a pre-glacial valley surface level in both the Severn and Marton valleys. But it has been so modified by glacial erosion and deposition that it is not possible to determine

the former direction of stream flow. In the Severn valley the platform is generally at a height of about 325 ft, and it apparently rises up the Marton valley to about 350 ft.

Both the Severn and Marton valleys are covered with glacial deposits, though some idea of the sub-drift surface is given by borehole records and exposures. At Abermule the Severn flow for some distance over bedrock on the north-westernside of the valley, at a height of about 300 ft. The opposite side is deeply covered with drift, and the bedrock surface is at about 270 ft at Abermule. About 700 yd downstream, at Bryn-Derwen Bridge the bedrock surface rises from about 270 ft in the south-east to 306 ft in the north-west. Bedrock must also lie at about 270 ft where the river cuts across the 325 ft platform at Bryn-llyn. The height of the sub-drift surface in the Severn valley below Montgomery Station is not known.

In the Marton valley, opposite Binweston Farm, the sub-drift surface is at about 270 ft. This is the only significant record for the Marton valley, and no bedrock exposures occur in the valley floor.

Since there has been considerable glacial erosion in both the Severn and Marton valleys, it is difficult to re-construct the height of the pre-glacial valley floor. The 325 ft platform in the Severn valley is probably the remains of an old valley floor, and the 270 ft sub-drift

on the south-eastern side of the valley may be the result of glacial overdeepening. Similarly, the 300-350 ft platform in the Marton valley is a pre-glacial valley level, and this has been overdeepened to about 270 ft opposite Binweston Farm.

It is difficult to envisage that the Marton valley was ever occupied by a through river. The configuration of the valley strongly suggests that the pre-glacial watershed was not far removed from its present position, in the neighbourhood of Marton village. The valley is narrowest at this point, and a prominent ridge, the surface of which is at 400 ft, occupies the centre of the valley. The watershed may have been temporarily destroyed while the ice front stood a little to the south-west, but it was restored by the drift accumulation.

It seems probable, therefore, that the pre-glacial course of the Severn was the same as at present, and that it did not flow either along the Marton valley, or through the Middletown gap.

# Chapter 9

## TOWARDS A CHRONOLOGY

### RELATIONSHIPS TO ICE MOVEMENTS IN OTHER AREAS

During the last glaciation the area under discussion was invaded by two major ice streams, moving in opposing directions from different sources. Drifts in the north-east are the product of an ice-sheet that originated as the outpouring of ice from southern Scotland and the Lake District. Moving across the floor of the present Irish Sea, it picked up marine shells and other material, in addition to its load of Scottish and Cumberland debris, and eventually impinged upon the coast of North Wales and Cheshire. One lobe of ice moved across Anglesey and down the west coast of Wales, while another pushed its way across the Cheshire and Shropshire Basin, probably as far south as Wolverhampton and Bridgenorth. The ice advanced against the northern slopes of the Long Mynd, to a height of at least 1250 ft., and penetrated southwards along the Church Stretton valley,

> " ... channeled in this direction by the barrier of the Long Mynd to the west, and, to a lesser degree, by Wenlock Edge in the east."[1]

The ice advances as far as Marshbrook in the Stretton valley, Rushbury in Ape Dale[2], and overrode the interfluves up to about 1,000 ft.;

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1. Mackney, D. & Burnham, C.P., 1966, p. 16.
2. Greig, D.C., and others, 1968, p. 295.

Hope Dale, Wenlock Edge and the Aymestry Ridge were all surmounted. (Fig. 9.1). To the south-west of Bourton the boulder clay reaches a height of about 900 ft. on top of the Aymestry Ridge. At Bourton (Burton) Mackintosh described,

> "... a remarkable load of Eskdale granite and Lake District felstone boulders..."[1]

In Corve Dale the ice penetrated about 1 mile south-west of Brockton.

The remainder of the area is dominated by drifts of a Welsh type, the product of ice streams which advanced into the region from the mountains of Central Wales. One stream passed over the Clun Forest escarpment, through a col at Black Gate at the head of the Mule, at a height of 1,368 ft. Another stream moved north-eastwards along the Severn valley towards Welshpool. Near Llandrinio it met a great efflux of ice moving down the Tanat and Vyrnwy valleys, and the combined streams passed south-eastwards towards Minsterley, Westerley and Church Pulverbatch. The Welsh and Northern ice sheets were confluent on the eastern side of the Habberly valley, against the high ground between Longden Manor and Cothercott Hill. The Northern ice was prevented from moving further westwards, but the Welsh ice advanced further east, producing intermingling of the drifts. A third ice stream moved into the area from the Clun valley, and pushed eastwards and northeastwards into lower Ape Dale and along the Marshbrook-Horderley valley towards Little Stretton where it was confluent with Northern ice. The Welsh ice deposited till, erratics and eroded meltwater channels up to at least 1,300 ft.

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1. Mackintosh, D.C., 1879, p. 442.

in this part of the Welsh Borderlands.

Irish Sea Ice

High-level drifts deposited by the Irish Sea ice during the last glaciation are common on the outer margins of Wales, the western flanks of the Pennines, and on the Isle of Man, at heights between 900 ft. and 1,400 ft. Perhaps the most famous are those on the summit of Noel Tryfaen in North Wales, first described by Trimmer[1] in 1831. False-bedded, calcareous sands and gravels, closely resembling beach deposits, containing chalk flints and Eskdale granites and numerous shell fragments, were found at a height of about 1,350 ft., and as low as 1,170 ft. Mackintosh compiled a list of the marine shells. These included Cardium echinatum, Cardium edule, Cyprina islandica, Astarte borealis, Tellina balthica, Mactra solida, var. eliptica and Saxicova rugosa.[2] On a recent visit to Moel Tryfaen[3], up to 20 ft. of sands, with gravel lenses, were found to lie between two Welsh tills. The overlying, non-calcareous till contained no Northern erratics and was 6-8 ft. thick, whilst the underlying till was up to 40 ft. thick. It would appear that the two ice masses were here contemporaneous, and that an interdigitation of their deposits occurred.

Mackintosh[4] described similar shelly gravels on the flanks of Halkin Mountain, at heights of between 750-900 ft. The gravels occur as kames and contain chiefly local limestone and grit of Carboniferous

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1. Trimmer, J., 1831, p. 331-332.
2. Mackintosh, D. 1881, p. 356.
3. The author, together with I.D. Ellis-Gruffydd and H.D. Foster, visited the 'classic' site in 1966.
4. Mackintosh, D., 1881, p. 357-359.

age, but also some Eskdale granites and numerous shell fragments. The most striking mound is Moel-y-Crio, at 950 ft.

On the eastern flanks of Esclusham and Ruabon Mountains, between Minera and the Vale of Llangollen, numerous mounds composed of gravels and stratified sands, with Northern erratics and shells, are found at heights of between 1,000 ft. and 1,350 ft.[1] To the west, at higher elevations, the Northern drifts give way to till and erratics of Welsh origin, including numerous felsites,

> " ... from the volcanic regions of Arenig and possibly Snowdonia."[2]

The boundary between the Welsh and Irish Sea drifts runs in a line from the Nant-y-ffrith valley to Bwlchgwyn and thence to Minera; the Cefn-y-fedw Sandstone escarpment of Esclusham and Ruabon Mountains to Garth Trevor, and thence along the course of Offa's Dyke to Chirch Castle, Castle Mill and Bron-y-garth.(fig. 9.1).

> "Where the Lower Carboniferous Rocks and the Cefn-y-fedw Sandstone have made a conspicuous escarpment the boundary is sharp, as on Esclusham and Ruabon Mountains, where Arenig boulders occur on the lower escarpment of the limestone but fail to cross the higher one of the Cefn-y-fedw Sandstone. South of the Dee valley no such barrier exists, and there is some intermingling of the drifts.[3]

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1. Mackintosh, D., 1882, p. 186-190.
2. Wedd, C.B., Smith, B., & Wills, J.L., 1928, p. 137.
3. Ibid., p. 138.

On the mountains to the west of Oswestry, Irish Sea drifts occur up to 1,120 ft. in the Selattyn district and 1,167 ft. on top of the Cefn-y-fedw escarpment.[1] Nicholson[2] has described these drifts in some detail, including the large gravel ridge at Gloppa, 2 miles north-east of Oswestry. The deposit ranges from 900 ft. to 1,160 ft., but the main mass is from 1,000 ft. to 1,150 ft. About 60 ft. of gravels were exposed in a pit at Gloppa and contained, in addition to local erratics, Ennerdale, Eskdale and Criffel granites, chalk flints and marine shells. The latter are

> " ... much broken, rolled and striated, some very fragmentary, but the bulk are in fairly good condition."[3]

Nicholson compiled a faunal assemblage of the shells, and was of the opinion that:

> "These gravels may probably be correlated with the Moel Tryfaen deposit, which contains a similar fauna of much the same general character, together with striated erratics and boulders."[4]

Kendall later revisited the section and saw, in the shelly gravels, intercalations of a greenish clay with exclusively Welsh rocks. This suggested to him that the ice streams from two quarters were here

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1. Wedd, C.B., Smith B., King W.R., and Wray D., 1929, p. 156-160.
2. Nicholson, A.C., 1892, p. 86-90.
3. Ibid., p. 89-90.
4. Ibid., p. 89.

contemporary, and that each in turn gained an occasional mastery.[1] The Irish Sea ice seems never to have invaded the western hills, and no trace of its presence has been found west of the Carboniferous escarpment that ranges southwards through Selattyn Hill to Mynydd-y-myfry, Sweeney Mountain and Llanymynech hill.[2] West of this line the area was occupied by Welsh ice.

> "As in Denbighshire, the elevated outcrop of the Cefn-y-fedw Sandstone, by the failure of northern boulders on it and of red boulder clay that contains these and fragments of marine shells, marks the western limit of the Northern Ice-sheet."[3]

On the opposite side of the Shropshire-Cheshire Basin, Irish Sea drifts of a similar nature occur at high levels on the flanks of the Pennines. Mackintosh described mounds of sands and gravels with northern erratics and marine shells at an altitude between 1,100 ft. and 1,300 ft. near Setter Dog Inn, Macclesfield.[4]

The similarity of these high level drifts, which may be traced for long distances in the Welsh Borderlands, the similarity of the faunal assemblages of the included marine shells, and the fact that the gravels all occur at roughly the same altitude on the outer slopes of the mountain districts, suggests that they are referable to the same glacial episode. The Welsh and Irish Sea ice masses were roughly

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1. Kendall, P.F., in discussion, Dwerryhouse, A.R., & A.A. Miller, 1930, p. 128.
2. Wedd, C.B., et. al. 1929, p. 156.
3. Ibid. p. 168.
4. Mackintosh, D., 1881, p. 363.

contemporaneous, and at Moel Tryfaen and Gloppa their drifts are interdigitated. The drifts on the northern flanks of the Long Mynd are similar in many respects to those described above, and the Welsh and Irish Sea ice masses were evidently contemporaneous, as in other areas. It thus seems reasonable to correlate the high-level drifts of the Long Mynd with those of Gloppa, Denbighshire, and Moel Tryfaen.

The occurrence of marine shells within the drifts led Mackintosh to believe there was a widespread submergence of the Welsh Borderlands, and that the erratics were transported by floating ice.[1] In this context, it is interesting to note that he also believed that the deep valleys of the Long Mynd are due to marine action,

> " ... excavated while the mountain, as an island, was gradually, or at intervals, suddenly rising above the sea."[2]

The Welsh drifts he thought was deposited by terrestial glaciers. It is now generally accepted that the high-level shelly drifts are the product of an ice sheet which advanced to the margins of Wales from the Irish Sea basin. In the Wrexham area,

> " ... the lobe of ice ... was at least 1,500 ft. thick at its centre. For, pushing up the ancient Dee valley (now below sea level), it mounted the dip slope of the Lower Carboniferous outcrop ... to heights of 1,000 to 1,250 ft. above sea-level, leaving

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1. Mackintosh, D., 1879, 1881, 1882.
2. Mackintosh, D., 1866, p. 159.

> there fragments of marine shells swept up from the floor of the Irish Sea."[1]

In Shropshire, the Irish Sea ice was probably of a similar thickness, for borehole records show parts of the Plain to be below sea-level.[2]

Welsh Ice

Another major ice stream moved into the area from the mountains of Central Wales, passing down the Severn valley and over the watershed of the Clun Forest escarpment. To the south, the easterly movement of ice into the district around Clun is indicated by the dispersal of large erratic boulders, including Rhayader grit and Llandovery grits and shales (fig. 6.2). They are most plentiful on the top of the ridge of hills, but are by no means confined to high levels. The highest boulder, a grit from Rhayader is on Black Hill at over 1,400 ft. Lapworth considered that these erratics belong,

> " ... to the Plinlimon group of Central Wales. The hills from which they have been derived are all south of Bala, and situated almost due west from their present position."[3]

The high ground of the Clun Forest escarpment, which runs southwards from Kerry Hill to Radnor Forest, is cut through by a number of cols, all of which appear to have carried ice from the west. In addition to the col at Black Gate, the watershed between the Teme and the Ithon is covered in places by considerable accumulations of drift,

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1. Wedd, C.B., et. al., 1928, p. 136.
2. Pocock, R.W., et. al., 1938, p. 186.
3. Lapworth, C., in Crosskey, H.W., 1882, p. 247.

extending up to 1,540 ft.,[1] and ice evidently passed over the watershed in an easterly direction. The basal layers of the ice moved southeastwards down the Teme and Clun valleys, while the upper layers went eastwards towards Clun. Ice also entered the head of the Lugg valley through a series of cols and an area of irregular ground between 900 and 1,200 ft. The access cols, at Bryngolfa (1,050 ft.), Maylord (925 ft.), Pen-y-clawdd (1,040 ft.) and Bleddfa (1,201 ft.) have all been modified to varying degrees by the passage of ice through them, and contain till deposits and ice-moulded landforms.[2]

The dominant feature in the glaciation of Radnor Forest was again the ice from Central Wales, which overrode it from the valleys of the Ithon and Wye. The ice apparently attained a higher level than further north, and erratics from the Ithon drainage have been found on the western face of the hill, up to a height of 1,750 ft. above Bwlchabryngolan.[3] Luckman found numerous erratics up to 2,000 ft. on the shoulder above 'The Riggles' near Shepherd's Well, including grits, greywackes and local Ludlovian rocks.[4]

> " ... the whole area between Radnor Forest and Pool Hill channelled ice into the head of the Lugg, at times overtopping both summits ... Erratics and possible subdued channel forms occur on Pool Hill itself, and the large 30 ft. channel which cuts right across Llanlluest at over 1,500 ft. clearly indicates the

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1. Dwerryhouse, A.R., and A.A. Miller, 1930, p. 103.
2. Luckman, B.H., 1966, p. 80-86.
3. Dwerryhouse, A.R., and A.A. Miller, 1930, p. 105.
4. Luckman, B.H., 1966, p. 87.

> depth and extent of ice using this major gap."[1]

The belt of high ground which forms the escarpment running south from Radnor Forest to the Wye gorge is cut by several passes, all of which carried the eastward. On the western slopes of Bryn-y-Maen, Dwerryhouse and Miller found numerous large boulders of dolerite at 1,400 ft., probably derived from Baxter's Bank, 9 miles to the north-west.[2] All along the escarpment morainic material is plentiful and many of the included erratics were derived from areas to the west.

It would appear, therefore, that the area between the Kerry Hills, Radnor Forest and the Wye Gorge was glaciated by ice derived from the west. Its source was probably in the Plynlimon massif, and the high ground extending from Llangurig in the north to Llanwrytyd Wells on the south. The upper layers of the ice evidently moved eastwards across Clun Forest, though the basal ice moved in directions which were for the most part conformable with the pre-existing relief.

In Cardiganshire, a westerly movement of ice from the Plyinlimon mountain area has been extablished:

> " ... between Plynlimon and the coast, the country was probably, at one time, completely covered by ice. The general direction of movement of this ice was towards the west from the high ground around Plynlimon itself. This is inferred from striations and the distribution of erratics of Plynlimon rocks."[3]

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1. Ibid.
2. Op. cit., p. 107.
3. Williams, K.E., 1927, p. 226.

Jones and Pugh also established a westerly movement of ice in the area south of Machynlleth which, with the exception of the lower slopes of the Dovey estuary, was glaciated by,

> " ... ice, (which) seems to have come from the high plateau region somewhere near, or to the east of Plynlimon and coinciding with the axis of the Central Welsh syncline."[1]

As in the borderlands, the Welsh ice was deflected from its course by southward-moving Irish Sea ice.

> " ... along the Cardiganshire coast, observations of striae shows that, here, the ice moved from north to south. It seems probable, therefore, that the southerly-moving Irish Sea ice acted as a barrier against the local Welsh ice, and that the latter was deflected into a north to south direction."[2]

Williams considered that the line of separation between the two ice sheets lay in the vicinity of Aberarth, and that to the south the Irish Sea ice impinged on the coast.

The distribution of Plyinlimon grits and other erratics from the Central Welsh Syncline, which occur both to the east and west of Plyinlimon, indicate that an ice-shed of a Central Welsh ice sheet must have lain across this mountain mass. It probably extended southwards over the high ground between Llangurig and Llanwrytyd Wells.

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1. Jones, O.T. & W.J. Pugh, 1935, p. 289.
2. Williams, K.E., 1927, p. 226.

That part of Wales lying west and north-west of Oswestry was glaciated by ice which originated not in the Plynlimon area, but in the Arenig and Aran Mountains. Kendall, in a summary of 'The Great Glacial Centres of Wales' stated that,

> " ... in North Wales, much admirable work has been done which clearly indicates the neighbourhood of Great Arenig (Arenig Mawr) as the radiant point for a great dispersal of blocks of volcanic rocks of a characteristic Welsh type."[1]

More recently, Foster has confirmed that a major ice cap developed in Merioniethshire and had a centre of dispersal trending approximately north-south, from west of the Arenigs to south of the Arans.[2]

In the area south of the Central Berwyn Axis and west of the Carboniferous escarpment in the Oswestry district, the general direction of ice movement was approximately E10°N, while the lower levels of the ice tended to follow the local trend of the valleys.[3] During the early stages of glaciation small glaciers originated in the Berwyn Hills. Later, ice from the Arenigs coalesced withthe ice-cap which had formed locally over the Berwyns,

> " ... forming a massive ice-sheet which completely buried the region, and in its progress overrode the hills and valleys without regard to their topography."[4]

Arenig felsites from the Merioniethshire ice cap have not been found

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1. Kendall, P. F., 1893.
2. Foster, H.D., 1968.
3. Wedd, C.B., et al., 1929, p. 158.
4. Travis, C.B., 1944, p. 16.

on the hills of the Welsh Borderland south of the Ceiriog valley. In the Tanat valley, the distribution of erratics of Llangynog felsite shows an eastward movement of local ice. Ice moving east and south-east from the Tanat and Vrynwy coalesced with that moving down the upper Severn and occupied the ground south of Oswestry. As it pushed outward the Welsh ice would shoulder off the Northern ice; the two ice sheets would combine and move in a resultant direction between south and east in the Shrewsbury area.

To the west of Wrexham the liberal scatter of erratic Arenig felsite boulders indicates a general direction of ice movement from west to east. Mackintosh[1] has traced these erratics in some detail in high and low level drifts, and showed that Welsh ice from Merionieth-shire extended down the Conway and Vale of Clwyd, and eastwards across the Clwydian Range as far as Halkin mountain.

> "Among the initial concentrations the one at the eastern end of Bala lake is most thickly strewn. A remarkable intermediate concentration occurs between Bryn Eglwys and Berwyn. Terminal concentrations ... are numerous in the neighbourhood of Llangollen; and there is a remarkable one around Eryrys, near Llanarmon."[2]

The distinctive erratics were found on the summit of the Eglwyseg escarpment at 1,600 ft. and at heights of 1,800 ft. and over on

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1. Mackintosh, D., 1879, 1873, 1874.
2. Mackintosh, D., 1879, p. 444.

Cyrn-y-Brain.

> "The greatest height I have yet found decided Arenig boulders is 1,900 ft.[1]

There is some evidence that the Welsh ice, in moving out of the Vale of Llangollen, overrode the ice from the north, and that the Welsh ice stream made its influence felt at a period when the northern ice had already established itself in the area.[2]

> " ... the main line of confluence of the two ice streams extended southward through Wrexham and Ellesmere towards Shrewsbury ... Along this junction we find, around Ellesmere and Cockshutt, a great development of morainic hills."[3]

Carvill Lewis considered this to be a medial moraine between the two ice sheets.[4] Another, apparently distinct, morainic belt, which extends east from Ellesmere to Welshampton, Whitchurch and to the north of Market Drayton, was deposited by Irish Sea ice during a pause in its retreat northwards.[5]

There is thus overwhelming evidence that there were at least two major centres of ice dispersal in Wales during the last glaciation - the Plynlimon ice cap in Central Wales and the Merionieth~~shire~~ ice cap in North Wales. From the distributions of erratics and inferred directions of ice movement it appears that much of Wales was completely buried by ice from these two centres. It is probable that the two major

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1. Ibid.
2. Pocock, R.W., & D. Wray, 1925, p. 65.
3. Ibid.
4. Lewis, H.C., 1894, p. 294.
5. Pocock. R.W., & D. Wray, 1925, p. 65.

centres of ice dispersal were merely constituent parts of a mountain ice cap, which had an ice shed now roughly followed by grid line Northing 82.

The Welsh ice masses were in contact with the Irish Sea ice along the North Welsh coast, in west Wales and in the Borderlands, where it extended up to at least 1,400 ft. Extraneous ice from the Irish Sea Basin was restricted to the peripheries of the Welsh massif. Only in the south did the Welsh ice have an outlet which was not blocked by Northern ice.

## GLACIAL CHRONOLOGY

Until recent years there existed a great confusion concerning the Welsh glacial chronology, particularly over the extent and age of the so-called Last Glaciation. This was largely because the dating of the drifts was based mainly on stratigraphical evidence, drift morphology, correlation with terrace sequences in extra-glacial areas, and presence or absence of periglacial structures.

Where glacial sands have been found lying between two boulder clays, this has often been taken to represent two distinct periods of glaciation separated by an interglacial phase. Since Hull's[1] recognition of a tripartite succession in the Pleistocene deposits of the Oldham area, the succession: Upper Boulder Clay, Middle Sands, Lower Boulder Clay has been accepted by many subsequent workers in the Shropshire-Cheshire Basin. The scheme has been largely adhered to by the Geological

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1. Hull, E., 1864.

Survey, particularly in the northern part of the basin. In the south, The Wem, Oswestry and Shrewsbury Memoirs do not refer to a tripartite division and attribute the deposits to a single glacial episode, with evidence of one minor fluctuation, the Welsh Readvance. Carvill Lewis was an early critic of the tripartite division, and regarded the drift succession as much more complex.

> " ... I have been far more impressed by the ever-varying deposits of the ice age than by any uniformity in their relations. Even in Lancashire, where the three-fold division was founded, the deposits are anything but constant in succession; and Mr. Binney of Manchester ... makes quite a different division from that of Professor Hull. Mr. Binney makes an Upper and Lower Sand and a middle Boulder Clay; While Professor Hull makes an Upper and Lower Boulder Clay and a Middle Sand."[1]

More recently, Poole and Whiteman[2] established the tripartite division in the Nantwich area, and extended it throughout the Shropshire-Cheshire Basin. They suggested that during the Lower Boulder Clay glaciation, the ice extended as far south as Birmingham, and that during its retreat deposited the Wolverhampton-Newport esker system, and the morainic belt at Ellesmere, both of which are of Middle Sands age. After the maximum limit of retreat had been reached ice sheets re-occupied the basin.

> "The sand and gravel ridges of the Middle Sands were but slightly modified during this

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1. Lewis, H.C., 1894, p. 70.
2. Poole, E.G., & A.J. Whiteman, 1961 and 1966.

> re-advance. This can be explained only if the deposits were waterlogged, as they are at present, and then frozen in a deep permafrost belt prior to the readvance."[1]

These ice sheets deposited the Upper Boulder Clay, which rests on the Middle Sands in the Ellesmere moraine, and which can be traced well-beyond the Shropshire-Cheshire basin, southeastwards. Thus the Ellesmere moraine, which was thought by Pocock to be a terminal feature of the Upper Boulder Clay ice, was considered by Poole and Whiteman to be a recessional feature of the Lower Boulder Clay glaciation.

It is becoming increasingly clear that the drift succession in the Shropshire-Cheshire Basin is often very complex, and the deposits cannot be simply divided into a tripartite sequence. Boreholes in the Shropshire area generally reveal a variable sequence of clays, gravels and sands which cannot easily be correlated with adjacent areas.[2] In some areas there are up to 5 clay bands separated by sands and gravels. Boulton and Worsley[3] considered that the concept of the Ellesmere moraine as Middle Sands overlain by Upper Boulder Clay to be a considerable oversimplification.

Serious doubt has been cast,

> " ... on the validity of the terms Upper Boulder Clay, Middle Sands and Lower Boulder Clay as meaningful lithostratigraphical units."[4]

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1. Op. cit., 1961, p. 120.
2. For example, the two Church Stretton boreholes, Appendix 1, p. 516
3. Boulton, G.S., & Worsley, P., 1965, p. 704.
4. Op. cit., p. 705.

and Thompson and Worsley state that a good deal of the Middle Sands are,

> " ... in reality, randomly distributed lens-shaped masses of fluvio-glacial sands within one layer of glacial deposits, representing varying environments of deposition occurring within a single stagnating ice sheet."[1]

In many areas, sands separating two boulder clays does not indicate two distinct glaciations; the sequence may be part of a complex series of deposits associated with a single ice-sheet.

Wills[2] correlated the deposits of the Shropshire area with terraces in the Severn valley. During the Main Irish Sea glaciation, the Irish Sea ice extended as far south as Wolverhampton and Bridgnorth, and the Welsh ice moved down the Severn valley towards Shrewsbury and another stream crossed the Clun and Radnor Forests. During the retreat of the Irish Sea ice, proglacial lakes Buildwas and Lapworth were formed, and the Iron Bridge gorge was cut as an overflow channel. The Marshbrook gorge was also initiated at this time.[3] The Main Terrace of the Severn, which is crowded with Irish Sea erratics, may be traced into the Iron Bridge gorge, and Wills correlated it with the Buildwas and Strethill Sands. The latter have been deposited close to the entrance to the gorge,

> " ... partly as marginal and partly as deltaic deposits,"[4]

during the retreat of the ice. The Main Terrace is essentially an outwash feature, deposited during the retreat of the ice from Iron

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1. Thompson, D., & P. Worsley, 1966, p. 198.
2. Wills, L.J., 1924, 1937.
3. Wills, L.J., 1924, p. 92.
4. Wills, L.J., 1924, p. 306.

Bridge, and it has been correlated with the No. 2 Terrace of the Avon into which it grades near Tewksbury.[1]

Wills correlated the Welsh Readvance with the Worcester Terrace of the Severn.

> "It is the highest terrace to be recognized on both sides of the Iron Bridge gorge, and its level there shows that the gorge was then some 60 feet shallower than now.[2]"

He also suggested the possibility that the Irish Sea ice sheet first withdrew completely and then re-invaded the northern part of the Shropshire-Cheshire basin, and deposited the Ellesmere moraine. The Welsh Readvance was not considered to be a major glacial period, since there is no evidence of an intervening interglacial.

Peake[3] is of the opinion that a similar re-advance occurred in the northern part of the basin. A tripartite division of the drifts was recognised. During the Lower Boulder Clay glaciation, which is correlated with the Main Irish Sea Glaciation, the ice penetrated as far south as Wolverhampton, reached mountain height along the Borderland, and carried northern erratics and marine shells to considerable elevations (fig. 9.1). The Middle Sands were deposited during a major retreat phase, and subsequently there was a readvance of Irish Sea ice, depositing the Upper Boulder Clay and the Ellesmere moraine. This 'Ellesmere Re-advance' of Peake is correlated with the Welsh Re-advance of Wills.

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1. Tomlinson, M.E., 1935.
2. Wills, L.J., 1937, p. 94.
3. Peake, D.S., 1961.

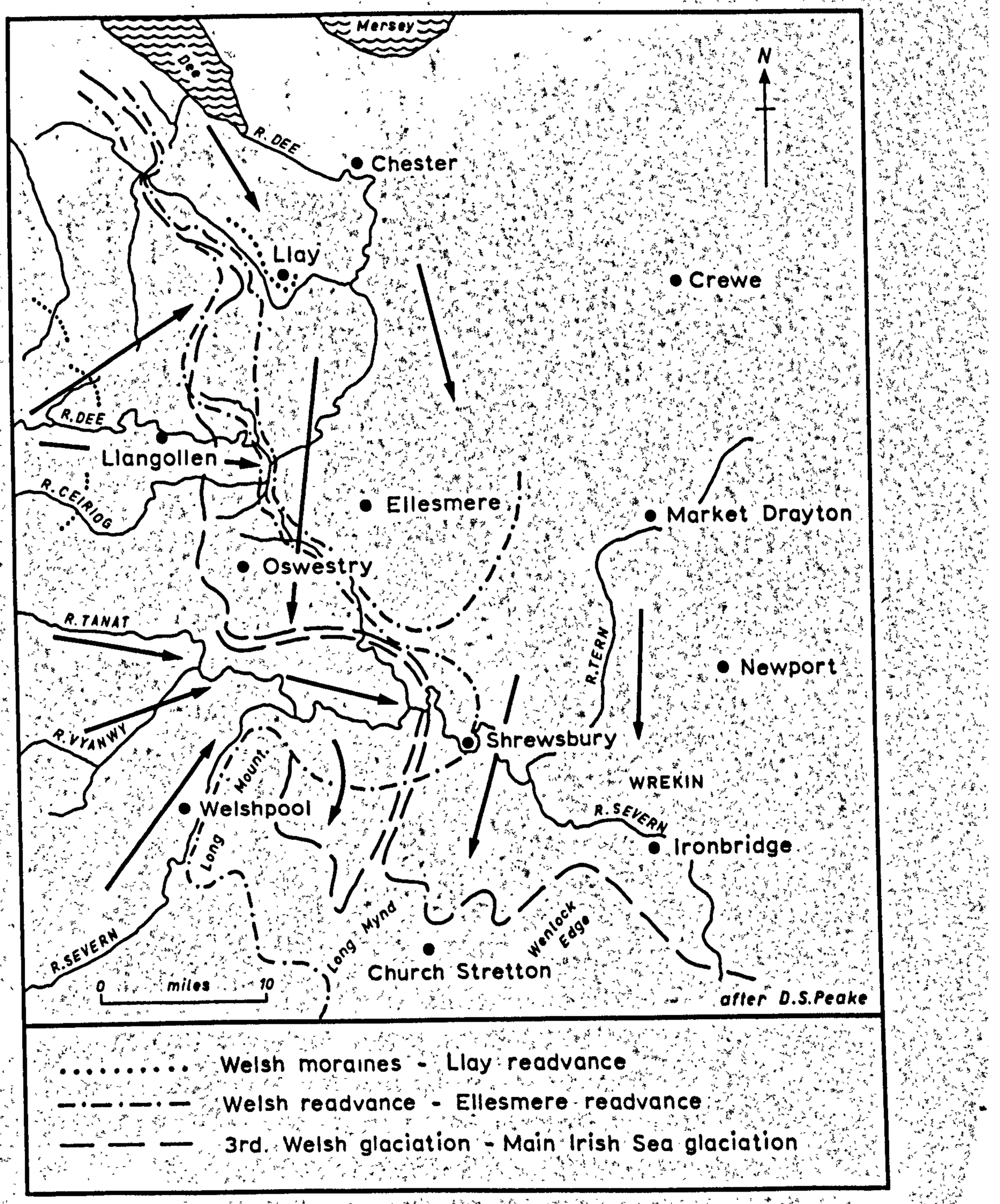


Fig. 9.1 Ice movements and glacial limits,

Outwash from the Ellesmere Re=advance forms the Wrexham Delta Terrace, and since this is overlain by till at Llay, it is considered that there was a further re-advance (Llay Re-advance), which is correlated with certain Welsh valley moraines, as at Pwll Glas.

Poole and Whiteman regarded their Upper Boulder Clay to be co-extensive with the Welsh Re-advance till, and proposed that the underlying Irish Sea gravels to be of Middle Sands age. In their view,

> " ... gravels which are older than the Welsh boulder clay have been correlated with gravels younger than it to form the basis of the Little Welsh Glaciation, and its correlation with the Worcester Terrace."[1]

They also refute Wills' correlation of the Ellesmere Moraine with the Welsh Re-advance; they regard the moraine as of Middle Sands age. It is concluded that,

> " ... the Little Welsh glaciation does not constitute a distinct episode."[2]

The Wrexham Delta Terrace is also regarded as Middle Sands.

It is abundantly clear that the attempts to base a glacial chronology on stratigraphical methods alone has led to considerable confusion in the Shropshire-Cheshire Basin. While Pocock and Peake regard the drifts to the south of the Ellesmere moraine as Lower Boulder Clay deposits, Poole and Whiteman traced the Upper Boulder Clay as far as Birmingham. The Main Irish Sea glaciation of Wills is equivalent to Peake's Lower Boulder Clay and to Poole and Whiteman's Upper Boulder

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1. Poole, E.G., & A.J. Whiteman, 1961, p. 107.
2. Op. cit., p. 108.

Clay. The latter find no evidence for a Welsh Re-advance, during which Peake believes the Ellesmere moraine was deposited. Pocock also maintained that the moraine was a terminal feature of the Upper Boulder Clay glaciation, though Poole and Whiteman regard it as a Middle Sands feature.

Pleistocene deposits are characterised by their extreme variability. The glacial deposits generally comprise complex sequences of tills, sands and gravels, subject to marked facies changes and containing a varied erratic suite. They can seldom be traced for long distances as well-defined stratigraphical or lithological units, and little is generally known of the actual conditions of deposition. Sands and gravels overlain by till does not necessarily mean that there has been a glacial re-advance.

Thus in establishing a Pleistocene chronology, the main requirement is for a dating method based on direct physical measurement and independent of the geological and biological context. Various methods of absolute dating have been developed in recent years, but radio-carbon dating, in particular, has become a useful tool for those who wish to date Pleistocene deposits in actual rather than relative terms. The author, together with H.D. Foster and I.D. Ellis-Gruffydd, visited a number of critical areas in Wales and the Borderlands in an attempt to clarify the glacial chronology.

## RADIO - CARBON DATING

### Buildwas

On both sides of the Severn valley, from Buildwas to the entrance

to the Iron Bridge gorge at Coalbrookdale, arelarge mound-shaped hillocks of sand and gravel, termed the Buildwas Sands by Wills[1] (fig. 9.2). The deposits om the lower slopes of the Tick Wood comprise a kame-complex with numerous enclosed depressions - Kettle holes - up to 15 ft deep. The upper limit of the deposits is about 300 ft, and they descend to the Severn valley at 150 ft. Those described by Maw[2] at Strethill, on the opposite side of the river, also rise up to 300 ft.

The stratigraphy of the Tick Wood deposits was examined at two large pits adjacent to Buildwas Abbey. In the pit to the east , the basal member is a yellowish-red (5YR 6/4), calcareous, false-bedded sand, in exsess of 60 ft thick. There are occasional coarser sand and gravel layers, coal horizons and small comminuted shell fragments. The base of the sands was not seen, and descends below the floor of the valley. There is a well-marked and sudden transition from the sands to the overlying well bedded coarse gravels, which are up to 12 ft thick. The gravels contain numerous shell fragments, in addition to a mixed Irish Sea and local Welsh erratic suite, the latter largely derived from the local Silurian outcrops. Towards the west the gravels become increasingly interbedded with coarse sands.

In the western pit about 40 ft of sands again form the base of the section, and there is a transition into the overlying bedded gravels and sands. The section is capped by up to 10 ft of ill-

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1. Wills, L.J., 1924, p. 277
2. Maw, G., 1864, p. 132-134

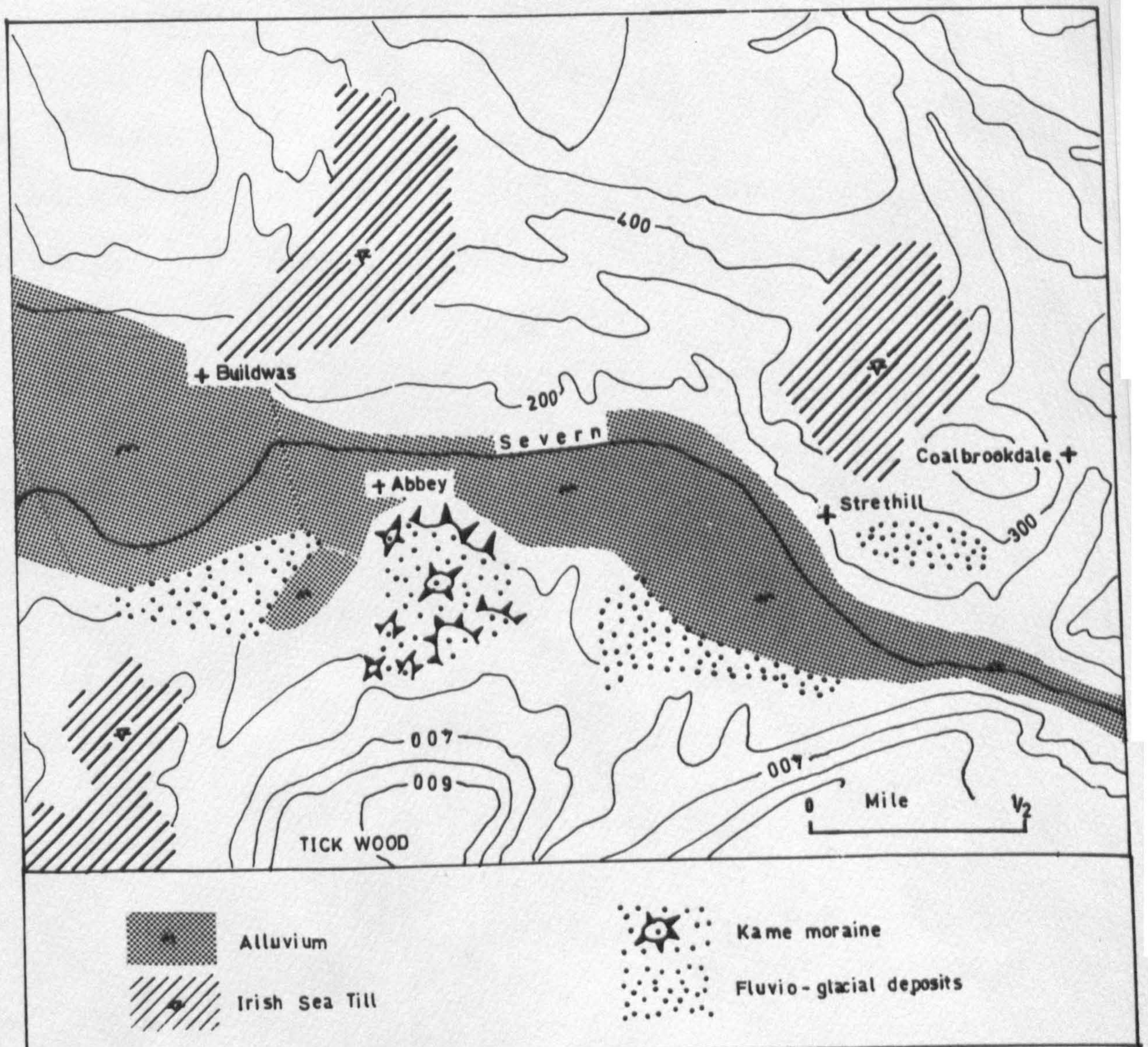


Fig. 9.2 Glacial deposits at Buildwas

sorted dirty gravels. The kame has a steep (30°) up-valley edge, which is probably an ice-contact slope, a concept supported by the interdigitation of the till and the down-valley dip of the sands and gravels.

Within the pit to the east, a mixed shell fauna was collected from the base of the gravels, about 12 ft. below the surface. The true depth below the original surface could not be accurately established since this has been stripped during recent quarry operations. A radio-carbon age determination on the mixed shell fauna yielded a date of $31,840 \pm \frac{1,660}{1,360}$ yr. B.P. (I-2939)[1].

New Marton

The deposits of the Bar Hill-Whitchurch-Wrexham moraine, or the Ellesmere moraine, are exposed in a kame complex at New Marton (SJ 335317). The succession recorded at this side was;

5. Up to 4 ft. of dark yellowish-brown (10YR 4/4), with till wedges.
4. Up to 6 ft. of reddish-brown (5YR 5/3) clayey till, with numerous erratics, some of which are rotted.
3. 12-15 ft. of well-bedded reddish-brown (5YR 5/4) sand, passes upward into inter-bedded sands and till.
2. 1-4 ft. of poorly-sorted gravels and pebbles, with a discontinuous horizon of organic silt.
1. ≠25 ft. of well-bedded and false-bedded reddish-brown (5YR 5/4) sands, with occasional lenses of

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1. All radio-carbon dating was carried out by Isotopes Inc., New Jersey.

fine gravel towards the top. A wedge of green stony till, about 1 ft. thick, was included, above and below which was a thin band of finely laminated red clay, up to 3 in. thick.

The erratic suite in the gravels and till was dominantly of 'Welsh' origin, including quartzites, Millstone grit, Coal Measure sandstone and some Arenig felsites. Occasional Lake District and Scottish igneous rocks were also found. A sample of the organic silt from the gravel horizon yielded a radio-carbon date of 26,270 $^{+\ 2,170}_{-\ 1,170}$ yr B.P. (I-2800). A preliminary examination of this silt revealed pollen of grasses, sedges and alder, which was considered typical of pool vegetation.[1]

Moel Tryfaen

On the summit of Moel Tryfaen (SH 518560), on the western margin of Snowdonia, a section towards the top of the Alexandra Slate Quarry revealed the following succession:

5. Slate tip
4. 6-8 ft. of grey, non-calcareous boulder clay, with abundant north Welsh erratics, often well-striated.
3. Up to 20 ft. of calcareous, false-bedded sands with gravel lenses. The sands strongly resembled a beach sand. The deposit contained abundant Irish Sea erratics and marine shells; in the sands the shells were generally highly comminuted, but larger fragments were found in the gravels.

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1. Davis, A., personal communication.

1. Up to 40 ft. of non-calcareous, grey till with abundant north Welsh erratics.

About 83 gm of marine shells and shell fragments were collected by H.D. Foster from the gravel horizon. Prior to dating, these shells were identified by Norton, who considered that they were

> "... of recent affinity (i.e. Post-Lower Pleistocene) and appear to represent beach and sub-littoral communities."[1]

The radio-carbon date of 33,740 $^{+\ 2,100}_{-\ 1,800}$ yr. B.P. (I-2803) agrees with the modern and boreal aspect of the fauna and its upper Pleistocene derivations, but as Norton stated,

> "... it would have been impossible to 'date' the assemblage by paleontological means."[2]

Mackintosh stated that,

> "... the finely laminated sands shows no trace of having ever been disturbed by the percolation of rainwater or frost, while the preservation of numerous shell fragments ... has evidently been owing to the clay preventing the downward passage of raihwater."[3]

It thus seems unlikely that the shells have been contaminated by the downward percolation of rain water, and the $C_{14}$ dates are not likely to be seriously in error.

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1. Norton, P.E.P., personal communication.
2. Ibid.
3. Mackintosh, D., 1881, p. 355.

Banc-y-Warren

On the west coast of Wales there is a well-developed kame-complex at Banc-y-Warren, Cardiganshire (SN 202482), previously dexcribed by Williams.[1] A large section is exposed in a gravel pit worked by the Cardiganshire Sand and Gravel Co. Up to 40 ft. of stratified sand is overlain by bedded gravels, pebbles and cobbles, passing upwards into more stratified sand, to a total thickness of about 110 ft.[2] Williams found that equivalent fluvio-glacial deposits nearby directly overlie stiff blue 'Lower Boulder Clay'. Welsh and Irish Sea erratics are common in the gravels, and shell fragments are scattered throughout the succession.

Some 70 ft. below the top of the section, within the gravel layer, are discontinuous horizons of peaty organic mud. A radio-carbon date of $31{,}800 \begin{smallmatrix} + 1{,}400 \\ - 1{,}200 \end{smallmatrix}$ yr. B.P. (I-2559) was obtained on a sample of the organic material. The point from which the sample was taken makes it unlikely that it has been contaminated by rainwater percolating downward.

Sources of Error

> "Radiocarbon dates are now very numerous and are used extensively, often with a high degree of faith proper to a precise scientific measurement and with insufficient appreciation of the possibility and magnitude of errors."[3]

Although radiocarbon assay is a very valuable tool in the absolute dating of Pleistocene events, it is subject to many sources

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1. Williams, K.E., 1927, p. 205.
2. The section and its implications are described in Brown, M.J.F. et. al., 1967 (a).
3. Shotton, F.W., 1967, p. 360.

of error which should be fully appreciated before any assessment of the significance of the date obtained can be made. Errors of three basic kinds reduce the value of the technique - statistical-mechanical errors, errors pertaining to the $C^{14}$ level of the sample itself, and errors related to laboratory treatment and measurement. The first is present as a result of the random disintegration of radioactive carbon, but is of no great concern as the practical sources of error inherent in the sample itself are far more serious.

An important assumption is that the concentration of radiocarbon in the atmosphere has remained constant in time. Some variation in the $C^{14}$ content is recognised, however, and it is thought that periods of higher concentration correspond with warm periods, and with the explosion in the atmosphere of nuclear bombs. Excesses of $C^{14}$ of as high as 8% have been recorded from woods of known age up to 5,000 yr. B.C.; this would correspond to an error of about 620 yr. This variation can be allowed for in the past 6,000 yr., and before that time it is thought that the error is proportionately small. It is doubtful if the error would be significant in dates of 20,000-40,000 yr. B.P. More important sources of error occur where there has been an exchange of radiocarbon since the death of the organism which incorporated it, leading to possibilities of contamination.

Perhaps the most serious problem is contamination by the intrusion of younger carbon into samples and,

> "...failure to recognise its presence has invalidated many so-called dates."[1]

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1. Op. cit., p. 362.

It can arise from the penetration of recent roots, humic acids, percolating groundwaters carrying down fresh calcium carbonate, and careless sampling procedures or laboratory techniques. Carbonate materials, especially marine shells, are unusually prone to contamination. This source of error is particularly significant in dating old samples near the limit of the technique. Shotton[1] estimates that for ages less than 20,000 yr., contamination must be very gross if they are to be greatly in error. However, a contamination of only 2% on a sample with a true infinite age could give a measured age of 32,500 yr. B.P.

Particular care was taken to minimise this source of error in sampling techniques. In all cases a fresh face in the quarry was exposed; the samples were taken up to 1 ft. behind the face, and as far as possible from the top of the section. Samples were collected with a clean hand trowel and immediately placed in clean plastic bags which were then sealed.

The dated shell material is perhaps the most suspect of the age-determinations presented. Percolating waters may not only cause a partial solution of carbonate, but may also foster an exchange between $C^{14}$ and $C^{12}$ with an enrichment in the former (fractionisation). These may lead to a serious underestimation of the true age of the sample. When using molluscs it is customary to remove a large part of the exterior of the shell by washing in dilute hydrochloric acid prior to dating. This is done in the laboratory, but the measured age may still be an underestimate. It is advisable to date only the largest shells, and these

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1. Ibid.

were systematically collected.

Contamination of the sample by old carbonaceous matter is a less serious source of error. Older carbon may be introduced in the form of fragmentary coal in sediments, or, as in the case of the Banc-y-Warren sample, by old (Tertiary) pollen spores in organic deposits. Mollusca add a certain proportion of ancient carbonates derived from old rocks to their tissue. In this way the radioactivity will be too low, and the indicated age too great. However, even when no steps are taken to remove the contaminant, 50% of dead carbon produces an overestimate of a half-life, 5,730 yr., irrespective of the true age. A more probable contamination of 10% represents an error of only 900 yr.

It would appear, therefore, that of the dates presented those for Buildwas and Moel Tryfaen, both of which were for shell samples, are the most suspect. Contamination by young carbonaceous matter is thought to be less likely at Moel Tryfaen because of the effective sealing of the sands by the overlying clays, preventing the downward percolation of water. At Buildwas contamination is more likely as the sands are not sealed, and the point from which the sample was taken was only 12 ft. from the top of the section.

Another source of error for all the dated samples lies in the fact that they are all essentially derived materials. There are mixed shell faunas, and the presence of Tertiary spores at Banc-y-Warren emphasises the fact that there are a variety of materials of different ages. Some is undoubtedly relatively old, but, as stated above, this need not be a serious source of error. Some organic matter will be

relatively young, possibly younger than about 30,000 yr. B.P., and this may lead to an underestimation of the true age of the deposit.

Radiocarbon dates are therefore 'absolute' only in a qualified way. An isolated date is to be regarded with caution. Despite the doubts and ambiguities that surround some of the dates presented, there is now a growing number of age-determinations for various sites in Wales and the Borderlands that, put together, tell a consistant story. In many cases the validity of the dates presented is supported by other, similar, assays in adjacent areas and even in neighbouring gravel pits. Many more dates are required, however, preferably at the same site and on different materials, to firmly establish a late Pleistocene glacial chronology.

Implications

The radio-carbon date of 31,840 yr. B.P. determined on the mixed shell fauna at Buildwas is significant in that the glaciation responsible for the deposition of the sands and gravels must post-date this period. The shells were picked up by ice from the floor of the Irish Sea basin, and are glacial erratics. Since the shells are a mixed fauna, some are probably older than the $C_{14}$ date, while others are probably younger. The date is important in that it gives a maximum age for the glaciation - post 31,840 yr. B.P. It has been established that the Buildwas Sands were deposited during the early stages of the retreat of the Main Irish Sea glacier, which is now seen to be equivalent to the Main Würm (Jung Würm) of Europe, which reached its maximum about 20,000 yr. B.P.[1]

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1. Gross, H., 1964, p. 187-197

(fig. 9.3). The marine shells were probably living during the Middle Würm (Mittlewürm), which in Europe extended from about 50,000 to 30,000 yr. B.P.

It was formerly believed that the Main Irish Sea glaciation was of Early Würm age, post-dating the Brørup Interstadial. At Upton Warren and Fladbury radio-carbon dates of 38,000 $\pm$ 700 yr. B.P. and 41,900 $\pm$ 800 yr. B.P. respectively have been obtained on organic material incorporated within the terraces of the Salwarpe and the Avon No. 2 Terrace.[1] (fig. 9.4). These have been correlated with the Main Terrace of the Severn, which was deposited as outwash during an early retreat phase of the Main Irish Sea glacier. More recently, Shotton[2] presented a further $C_{14}$ date of 28,200 $\pm$ 500 yr. B.P. from organic matter within the Avon No. 2 Terrace at Brandon, Warwickshire. The so-called 'Upton Warren Interstadial Complex' was therefore considered to extend from about 50,000 to 23,000 yr. B.P.

The dating of the drifts at Buildwas as of Early Würm age rested on two basic assumptions - that the correlation of the Main Terrace of the Severn with a retreat phase of the Irish Sea glacier is valid, and that the dated plant matter within the terraces is <u>in situ</u>. With reference to the first assumption, Hollingworth commented that,

> " ... the link in the chain of dating
> of the last ice ... was perhaps a weak one
> ... The correlation with the Main Terrace
> of the Severn and the Avon No. 2 Terrace

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1. Coope, G.R., Shotton, F.W. and I. Strachan, 1961, p. 381.
2. Shotton, F.W., 1967, p. 122.

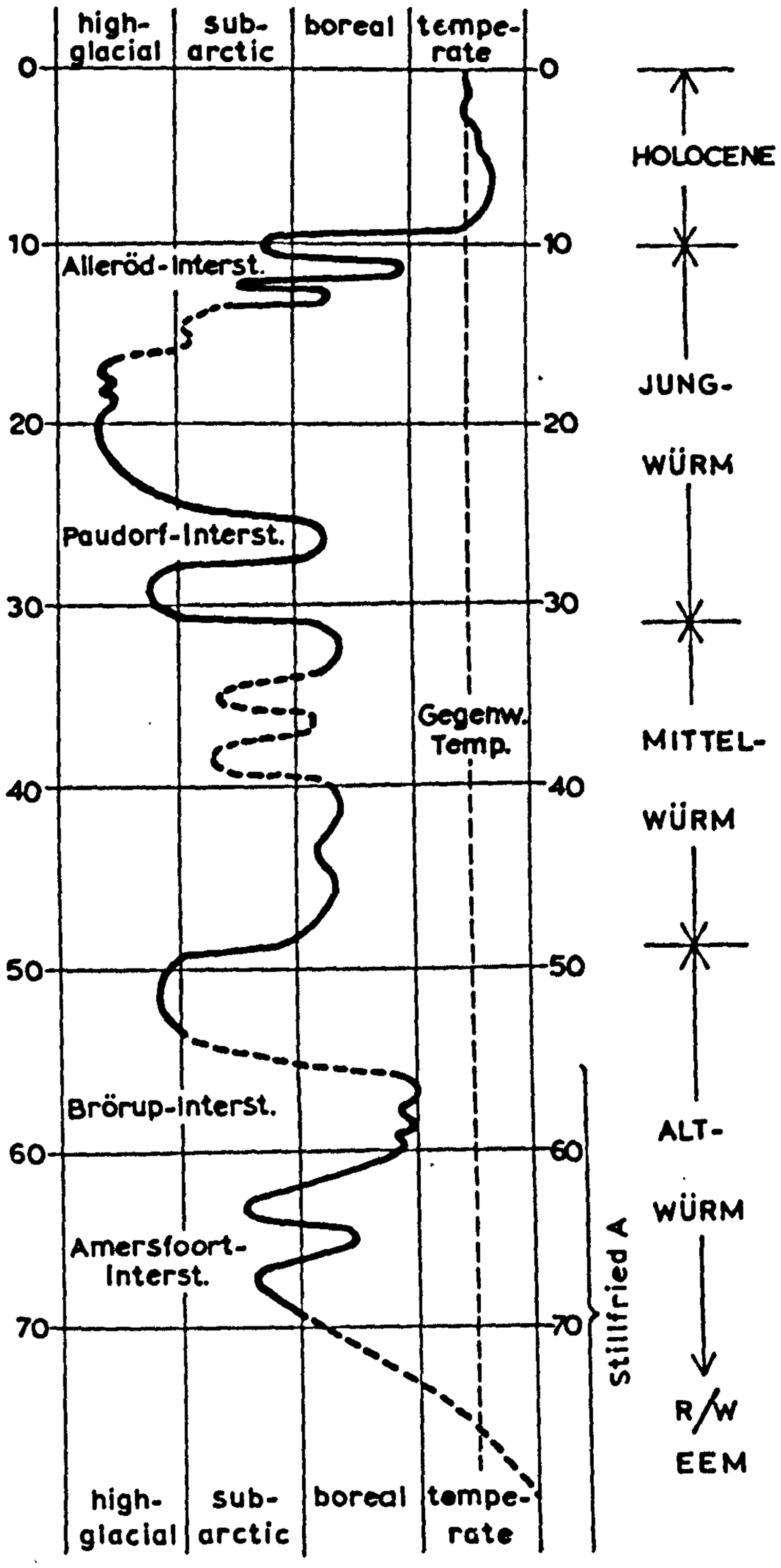


Fig. 9.3 Climatic record of the late Pleistocene

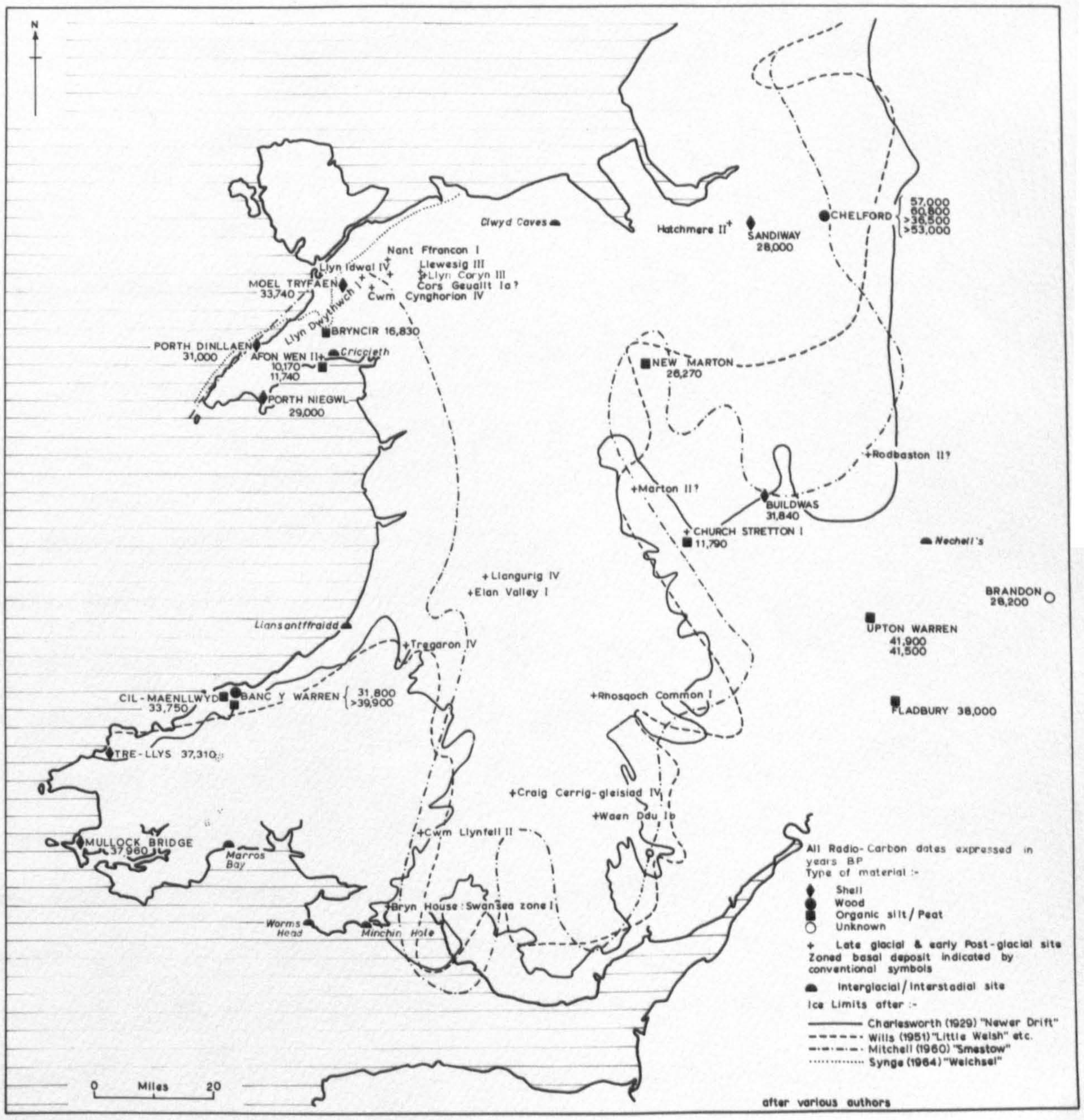


Fig. 9.4 Various radiocarbon dates and ice limits in Wales

> was via the Iron Bridge gorge, through which there would appear to be either erosional or depositional continuity to provide a direct link with outwash to the north. There was a possibility or probability that the Iron Bridge gorge could have operated as a direct glacial channel on two or more occasions."[1]

Gill also stated that,

> " ... it seemed that there might be something amiss with the correlation of the Severn terraces and the ice retreat stages."[2]

Wills, however, presented strong evidence, difficult to refute, to establish the correlation of the Main Terrace with the retreat of the Irish Sea glacier, and this is not doubted by the author. The second basic assumption must, therefore be examined.

Several lines of evidence suggest that the Upton Warren deposits are not *in situ*, but are derived. Coope *et al*.[3] recognised an extremely varied floral and faunal assemblage, which included dissociated bones of mammoth, woolly rhinoceras, associated with Arctis tundra conditions. In contrast, bones of frog and the three-spined stickleback (*Gasterasteus aculeatus*) were not uncommon, the former not occurring in full tundra conditions, while the latter,

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1. Hollingworth, S.E., in discussion. Shotton, F.W., & I. Strachan, 1959, p. 115.
2. Gill, W.D., *Ibid*.
3. Coope, G.R., Shotton, F.W., & Strachan, I., 1961, p. 244.

> " ... where the climate is severe
> is restricted to salt water."[1]

In contrast to the verterbrate fauna, the molluscs included abundant Pisidium vincentianum, the modern distribution of which is now Mediterranean.[2] The insect fauna included numerous beetles, of which 8 species live in the higher latitudes of Scandinavia; 13 species are north European, though still occurring in the north of the British Isles; whereas 20 species are relatively southern in their distribution, at least 15 of which have a present day European distribution which is almost entirely south of latitude 60° N.

The flora also demonstrates a variety of climatic conditions; while the greater part is temperate, there is a distinct boreal and arctic-alpine element.

Both the faunal and floral assemblages at Upton Warren contain elements that are geographically incompatible, and the explanation that the,

> " ... apparent climatic incompatibility
> of many of the species ..."[3]

represents,

> " ... a mingling of species character-
> istic of the earlief climate with the
> incoming species typical of the new
> conditions."[3]

at a time of rapidly ameliorating climate, is essentially catastrophic

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1. Ibid.
2. Dance, S.P., appendix to Coope, G.R., et al., 1961, p. 244.
3. Dance, S.P. op cit.

and leads to untenable ecological associations. For example, the virtual absence of trees and shrubs generally associated with a thermophilous ground flora, identified in the deposits, is explained by the grazing habits of large arctic-tundra herbivores, which can hardly have lived in temperate conditions, and the dissociated bones of which are almost certainly derived.

The mixed floral and faunal assemblages at Upton Warren are perhaps more readily explained if they are considered as being essentially derived, as are the terrace gravels in which they are incorporated. It is not doubted that the Upton Warren deposits are interstadial in character, and that they belong to the Middle Würm. If they are derived, the radiocarbon dates can only give a maximum age to the Irish Sea glaciation, post-dating 28,000 yr. B.P. This is in accord with the dating of the glaciation established at Buildwas.

Shotton has recently revised his views concerning the age of the Main Irish Sea glaciation.[1] In the Four Ashes area, sands and gravels, 4-15 ft. thick, are overlain by red Irish Sea till. Interbedded with the gravels are layers of organic silt and peat. At the base in hollows in the Bunter Sandstone, and at various levels within the body of the gravels. Two peaty samples, taken from different layers within the gravels, have been dated at 36,340 $^{+\ 770}_{-\ 700}$ yr. B.P. and 30,655 $^{+\ 765}_{-\ 700}$ yr. B.P. These dates indicate that the peat belongs to the Middle Wurm Interstadial, and it follows that the overlying till must be of Main Würm age. This casts further doubt on the validity of either the correlation of the Main Irish Sea glaciation and the Main Terrace of the Severn, or the in situ nature of the Upton Warren material.

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1. Shotton, F.W., 1967, p. 1366.

At Rodbaston, near Penkridge, nine miles north of Wolverhampton, a kettle hole formed during the retreat of the Irish Sea glacier contains sediments no older than Zone III or Zone II of the Late Glacial. On the assumption that the glacial deposits in this area were of Early Wurm age, Shotton and Strachan stated,

> " ... the problem remains unanswered as to why a hollow which came into existence on the deposits of the Irish Sea glacier should remain free of organic accumulation for about 30,000 yr. ... Difficult though it may be to picture and accept, the only explanation which seems to be available at present is that the ice mass of the future kettle hole remained unmelted over this long period."[1]

Hollingworth commented that he was also,

> " ... rather unwilling to envisage the hollow remaining empty for a period of 30,000 yr."[2]

and stated that,

> "Until the last ice in the Cheshire Plain was dated more directly, the possibility that the interval between its disappearance and the infilling of the Rodbaston kettle was a brief one should not be dismissed."[3]

Since the last glaciation of the area has now been dated as of Main

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1. Shotton, F.W., 1959, p. 5.
2. Hollingworth, S.E., in discussion, Shotton, F.W., & I. Strachan, 1959, p. 12.
3. Ibid.

Würm age, the once-anomolous time span between the disappearance of the ice and the infilling of the Rodbaston kettle hole no longer exists.

In the Vale of Stretton, Rowlands recorded an Allerŏd peat, dated at 11,790 $\pm$ 140 yr. B.P., overlying inorganic clays and gravel, presumably of Zone I age, which in turn rests on boulder clay.[1] He also found possible Zone III and Zone II deposits in the Marton valley, resting on boulder clay. Rowlands, however, agreed with Shotton's earlier view that the drifts of the Shropshire area are of Early Würm age. In which case there would be a considerable gap between the retreat of the ice and the deposition of the Zone I gravels. Such a time span no longer exists now that the drifts have been more firmly dated as Main Würm.

At New Marton, representatives of both the so-called Middle Sands and Upper Boulder Clay have been recorded. However, considerable doubt has been cast on these terms as meaningful lithostratigraphical units. Since the sands and till often interdigitate and numerous till wedges occur, it is probable that the deposits were laid down within a single stagnating ice sheet. The deposits of the New Marton area are referable to a single glacial episode. In view of the age-determination onthe organic silt, which is an integral part of the succession of well-sorted sands and gravels, this must have occurred later than 26,270 yr. B.P. Since the erratic content is preponderantly of Welsh origin, with a few Irish Sea indicators, a Welsh glaciation is implied. Further east, at Ellesmere, where the deposits form part of the same moraine,

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1. Rowlands, P.H., 1966, p. 103.

Irish Sea erratics are dominant. This suggests a contemporaneous Irish Sea and Welsh glaciation during the Main Wurm.

To the north of the Ellesmere moraine, at Sandiway, Boulton and Worsley obtained a date of 28,000 $^{+\ 1,500}_{-\ 1,800}$ yr. B.P. on two shells of Nucella lapillus from coarse sands at the base of a sand-till complex.[1] It seems logical that these deposits are contemporaneous with those at New Marton, and the date adds support to the view that the last glaciation of this area was of Main Würm age.

The succession exposed in the sand quarries of the Chelford area has been described by Simpson and West[2] as a type example of the tripartite drift sequence. It has been shown, however, that the drifts are rather more complicated, and that an Upper Boulder clay-sand complex overlies a series of strongly cross-stratified, poorly-sorted sands. These deposits are separated by a strong unconformity from a lower well-sorted sand sequence, the Chelford Sands.[3] Several lense-shaped organic beds occur within the Chelford Sands, and have been dated at 60,800 $\pm$ 1,500 yr. B.P.[4] which equates with the Brorup interstadial. The overlying glacial complex must, therefore, be younger than this, and since it is co=extensive with the deposits of Sandiway it must also belong to the Main Wurm glaciation.

Laterally analogous deposits of this same ice sheet appear to seal caves containing Aurignacian implements in the Vale of Clwyd.[5]

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1. Boulton, G.S., and P. Worsley, 1965, p. 706.
2. Simpson, I. and R. West, 1958, p. 239.
3. Boulton, G.S., and P. Worsley, 1965, p. 704-705.
4. Coope, G.R., 1959.
5. Hughes, T. McK., 1887, p. 73-120.

Movius concludes that the Aurignacian culture flourished in the Paudorf interstadial, immediately preceeding the Main Würm maximum.[1] This adds further evidence to the concept of a Main Würm glaciation in this area.

Boulton and Worsley were of the opinion that the Wrexham-Bar Hill-Whitchurch moraine marked the limit of the Main Würm glaciation in the Shropshire-Cheshire basin.[2] They based this view on differences in depth of carbonate leaching and morphological contrasts to the north and south of the moraine.

> "The boulder clays and related deposits to the south of the moraine are generally thinner and considerably more eroded than those to the north, often showing evidence of considerable solifluxion; depositional forms are rare."[3]

Depth of leaching is a function of several variables, including permeability, carbonate content, topography, climate and time; and although only a few measurements were made to the south of the moraine, compared with 54 to the north, the differences in depth of leaching were considered to be significantly different. It would seem, however, that many more measurements need to be taken south of the moraine before this can be established.

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1. Movius, H.L., 1960, p. 355-391.
2. Boulton, G.S., & P. Worsley, 1965, p. 705-706.
3. Ibid.

Moreover, the morphological contrasts north and south of the moraine are not as strong as Boulton and Worsely suggest. Harrison, with reference to the drifts of the Midlands, stated that,

> " ... the freshness of the deposits and the small amount of denudation to which they have been subjected, shows that no long period of time (probably not more than 10,000 years) has elapsed since the close of the glacial period."[1]

The drifts often attain considerable thicknesses, often in excess of 100 ft., as near Shrewsbury, the Vale of Stretton, and the Mule valley.

Considerable doubt is therefore cast on the evidence used by Boulton and Worsely to determine the age-difference of the drifts to the north and south of the Ellesmere moraine. Radio-carbon dates in the Buildwas area and at Four Ashes indicate that during the Main Wurm glaciation the Irish Sea ice advances as far south as the Wolverhampton-Church Stretton-Bridgenorth line. If there is a significant difference in the depth of leaching of the deposits it is probably because the drifts to the south of the moraine were uncovered by ice at an earlier stage of deglaciation than those to the north. It is likely that the Ellesmere morainic complex was laid down during a pause in the retreat of the Irish Sea and Welsh glaciers.

The radio-carbon date determined on the mixed shell fauna from within sands and gravels on the summit of Moel Tryfaen indicates that the glaciation repponsible for their deposition post-dates 33,740 yr. B.P.,

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1. Harrison, W.J., 1897.

and is of Main Wurm age.

Baden-Powell[1] argues that prior to the last glaciation, the isothermal gradient of the eastern Atlantic was much steeper than today, and this would permit a mixed fauna, such as was found on Moel Tryfaen, to live in close proximity. Theoretically, the members of the fauna could date from any time within the period between the retreat of the immediately preceeding glaciation and the glaciation in which the fauna now occurs; representatives of the Eemian (Riss-Wurm) interglacial and the Wurm interstadials might be expected. Contamination by old carbonaceous matter is likely to affect the $C_{14}$ age-determination, and will result in 'ages' that are overestimated. Shotton states, however, that,

> "This is not...a serious source of error even when no steps are taken to remove the contaminatant: 50 per cent of dead carbon produces an overestimate of a half-life, 5730 years, irrespective of true age; a more probable contamination of 10 per cent represents an error of only 900 years."[2]

Contamination by modern carbon is a more serious problem, and can arise from the penetration of rootes or downward percolation of rain-waters. However, this is not likely to have seriously affected the Moel Tryfaen date. The shells were collected some 20 ft. from the top of a fresh section; there were no roots or root-channels present, and the presence

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1. Baden-Powell, D.F., 1956.
2. Shotton, F.W., 1966, p. 361-362.

of clay overlying the sands and gravels would effectively prevent the downward percolation of rain water. The age determined on the shells is thus considered significant, and during the Main Wurm glaciation the Irish Sea ice extended to the summit of Moel Tryfaen. Since the sands and gravels lie between two Welsh tills, it is probable that the Welsh and Irish Sea glaciations were contemporaneous in this area.

The deposits on the summit of Moel Tryfaen have been correlated with other high-level shelly drifts on the Welsh Borderland, particularly with those at Gloppa, near Oswestry. At Gloppa, Lapworth considered that the interdigitation of the Welsh and Irish Sea drifts suggests the contemporaneity of the two ice sheets. The Gloppa deposits are also preferable to a Main Wurm glaciation, as are the high-level drifts on the northern slopes of the Long Mynd. The height of the drifts at Gloppa also makes it extremely unlikely that the Ellesmere moraine represents the limit of this glaciation.

The radio-carbon date of 31,800 yr. B.P. obtained on organic silt from within sands and gravels at Banc-y-Warren, Cardiganshire, suggests that the glaciation responsible for their deposition is of Main Wurm age. In a neighbouring pit, at Cil-maenllwyd, wood fragments, collected by John form Sands and gravels with similar stratigraphical relationships, were dated at 33,750 $^{+\,2,500}_{-\,1,900}$ yr. B.P.[1] This date agrees closely with that obtained at Banc-y-Warren.

Boulton criticised John's age determination of the grounds that the pollen analysis indicated a floral assemblage which was unlikely to have existed at 33,000 yr. B.P.[2] Some elements must date from a relatively warm interstadial, older than the Middle Wurm, or from an

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1. John, B.S., 1967, p. 630-633.
2. Boulton, G.S., 1968, p. 190-191.

earlier interglacial period. Indeed, John mentions the presence of Carboniferous and Mesozoic spores and pollen in the sample, which emphasises the fact that the remains have come from a variety of deposits of different ages. However, as stated above, contamination by old carbonaceous matter is not likely to cause a serious error in the age-determination. Contamination by young matter is more likely to introduce error. John considers that contamination of the wood by percolating rain-water is unlikely. Boulton's arguement that since the organic matter is derived there will have been some contamination by organic matter younger than 33,000 yrl B.P. is more probable. However, only the wood fragments, and not other organic matter, was dated, and it is likely that they are of similar age and unlikely to be seriously affected by younger organic matter. It would seem, therefore, that John's date is valid and that, together with the date of 31,800 yr. B.P., it is strong evidence for a Main Wurm glaciation in this area.

Further south, along the Pembrokeshire and Cardiganshire coast, John[1] obtained two radio-carbon dated of $37,960 \begin{smallmatrix} + 1,700 \\ - 1,400 \end{smallmatrix}$ yr. B.P. and $37,310 \begin{smallmatrix} + 1,515 \\ - 1,275 \end{smallmatrix}$ yr. B.P. from a mixed shelly fauna collected from within glacial sands and gravels at Mullock Bridge and Tre-llys. John used the dates in support of a great deal of other evidence from West Wales to propose a Middle Wurm interstadial and a Main Wurm glaciation in the area.[2]

Williams[3] has shown that the Irish Sea and Welsh glaciations were contemporaneous in west Wales. He considered that the line of separation

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1. John, B.S., 1965 (a), p. 622-623.
2. John, B.S., 1965 (b).
3. Williams, K.E., 1927.

between the two ice sheets lay in the vicinity of Aber-arth, and that to the south the Irish Sea ice impinged on the coast. The westward movement of ice from the Plynlimon massif, established by Jones and Pugh[1], must also, therefore, have been of Main Wurm age.

There is thus considerable evidence to suggest that during the Main Wurm the Irish Sea ice extended at least as far south as Milford Haven in West Wales. This is at variance with the published views of Mitchell[2] and Synge[3], who considered that the Bryncir-Clynnog Fawr moraine on the Lleyn peninsula marked the limit of the Main Wurm glaciation, and that all drifts to the south are at least of Early Wurm age. The author, together with Foster and Ellis-Gruffydd[4] have suggested that the Bryncir moraine is, in fact, the limit of the Scottish Re-advance into North Wales.

Exposed in a small quarry cut into one of the elongated mounds of the moraine is a complex series of false-bedded sands and gravels, up to 35 ft. thick and dipping in a southerly direction. The erratic sui te of this area is mainly Snowdonian, mixed with some Irish Sea material. Isolated nodules of organic silt are scattered throughout the entire section, and a sample collected 20 ft. from the top of the section was subjected to $C_{14}$ dating. The sample has been dated as $16{,}830 \begin{smallmatrix} + 970 \\ - 860 \end{smallmatrix}$ yr. B.P. (I-2801). This suggests that the glaciation responsible for the deposition of the moraine occurred after this period. The dominance of

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1. Jones, O.T. and W.J. Pugh.
2. Mitchell, G.F., 1960,p. 313.
3. Synge, F.M., 1960, p. 15.
4. Brown, M.J.F., H.D. Foster, & I.D. Ellis-Gruffydd, 1967, (b).

Snowdonian erratics implies a movement from Snowdonia, as well as the Irish Sea, at this time.

Contributary evidence establishing the validity of the date of this glaciation is found at Afon Wen. Here, kettle holes developed within the glacial outwash gravels from the Bryncir moraine have been found to contain Zone II peats, underlain by Zone I grey-brown clay. The Allerod peat is overlain by solifluxion gravels, referred to Zone III.[1] Saunders obtained a $C_{14}$ date of 11,740 $\pm$ 170 yr. B.P. on peat collected from a kettle hole developed in the outwash of this ice, at Glanllynau.[2] Saunders also agrees that the Bryncir moraine marks the limit of a Scottish Readvance during early Zone I times.

Penny attributed,

> " ... the Scottish Re-advance and the Perth-Aberdeen Re-advance to Zone I, and since Zone I is known to contain two cold periods (Ia and Ic) separated by the Bølling Oscillation, it is tempting to suggest that the Scottish Re-advance belongs to Zone Ia and the Perth-Aberdeen Re-advance to Zone Ic."[3]

but stated that,

> " ... there is not terminal moraine marking the limit of this Scottish Re-advance...(it)... is merely the mapped feather edge of the Upper

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1. K. Simpkin, personal communication.
2. Saunders, G.E., 1968, p. 77.
3. Penny, L.F., 1964, p. 399-400.

Boulder clay."[1]

It is suggested that the Bryncir-Clynnog Fawr moraine forms part of the missing terminal moraine of the Scottish Re-advance.

At Dinas Dinlle, on the north east coast of the Lleyn Peninsula, a complex drift sequence is exposed in cliffs up to 100 ft. high. (SH 4-5). The stratigraphy of this section was described by Jehu,[2] and later by Synge,[3] who considered that the deposits are of two distinct ages. The Middle sands and gravels were found to be heavily iron-stained in the upper horizons and well-developed iron-pans are displaced by faulting. At the southern end of the section a distinct weathering horizon, some 2 ft. in depth, occurs at the top of the gravels, which, together with the frequent induration and frost structures, are evidence of a distinct stratigraphical break between the sands and gravels and the overlying 'Upper stony clay'. The latter showed far less disturbance and was not indurated or iron-stained. Synge has shown that the upper stony clay is syncronous with the Bryncir moraine.

Ellis-Gruffydd[4] was first to recognise that the whole sequence at Dinas Dinlle has been faulted and folded, and suggested that these are glacio-tectonic structures. They include a well-developed anticline disturbed by overthrust and reverse faulting, giving way to gentler folding at the southern end of the section. They are clear evidence that the deposits have been overridden by ice from the north-east, probably by Scottish Re-advance ice.

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1. Ibid.
2. Jehu, T.J., 1909.
3. Synge, F.M., 1964.
4. Brown, M.J.F., Foster, H.D. & I.D. Ellis-Gruffydd, 1967 (b).

In conclusion, there is overwhelming evidence that during the Main Wurm the Merioniethshire and Central Welsh ice caps were active, and that these Welsh ice masses were in contact with Irish Sea ice along the north Welsh coast, in west Wales and the Borderlands. It is probable that much of Wales was buried by ice during the Last Glaciation, but the Irish Sea ice was restricted to the peripheries of the Welsh massif. The position of the 'Welsh Re-advance' is not clear, and the evidence in the area south of Oswestry is not conclusive that a major re-advance of ice occurred. There may have been a minor fluctuation of the Welsh ice sheet after the partial withdrawal of the Irish Sea ice. It is fairly clear, however, that a re-advance of Welsh and Irish Sea ice occurred in the Lleyn Peninsula during Zone I times.

Much of the dated organic material falls within the Middle Wurm interstadial (Upton Warren interstadial), during which there is still a large measure of uncertainty concerning climatic fluctuations. It is also likely that there were minor climatic fluctuations during the retreat of the Main Wurm ice sheets, but these still need to be worked out in greater detail. It is doubtful if the retreat was continuous, and there may well have been a major pause, during which the Ellesmere moraine was deposited.

# APPENDIX I

## BOREHOLE DATA

Except where indicated, all records were consulted at the Institute of Geological Sciences, London.

Sheet SO 19

| | Thickness | Depth (ft) |
|---|---|---|
| KERRY (143901)[1] | | |
| Topsoil - loam | 2-0 | 2-0 |
| Loam - Clay | 5-0 | 7-0 |
| Fine gravel | 9-0 | 16-0 |
| Sandy clay | 7-0 | 23-0 |
| Clay | 13-0 | 36-0 |
| Alluvial grit and sand | 64-0 | 100-0 |
| LLANDYSSLE (195954) | | |
| Soil | 2-0 | 2-0 |
| Gritty marl | 36-6 | 38-6 |
| Light coloured shale | 4-0 | 42-6 |
| Marl and gravel | 4-6 | 47-0 |
| Blue shale | 40-0 | 87-0 |
| ABERMULE (163945) | | |
| Soil | 4-0 | 4-0 |
| Marl and gravel | 43-0 | 47-0 |
| Blue rock | 32-0 | 79-0 |
| Granite | 20-0 | 99-0 |
| Blue marl and rock | 17-0 | 116-0 |
| Granite and marl veins | 4-0 | 120-0 |
| ABERMULE (164945) | | |
| Soil | 3-0 | 3-0 |

1. Grid References on 1:25,000 sheets.

| | Thickness | Depth |
|---|---|---|
| Stony marl | 3-0 | 3-0 |
| Sand | 1-6 | 7-6 |
| Marl and gravel | 17-6 | 25-0 |

ABERMULE - THE COURT

| | | |
|---|---|---|
| Soil | 2-0 | 2-0 |
| Marl | 7-0 | 9-0 |
| Dry sand and gravel | 2-0 | 11-0 |
| Stony marl | 10-0 | 21-0 |
| Gravel | 8-0 | 29-0 |
| Shale rock | 3-0 | 32-0 |

ABERMULE - BRYN-DERWEN BRIDGE

A. 313.41 ft. (163951)

| | | |
|---|---|---|
| Brown clay | 6-6 | 6-6 |
| Brown and grey shale, severely shattered | 14-6 | 21-0 |
| Grey shale, shattered | 37-0 | 58-0 |

B. 310.1 ft. (16309505)

| | | |
|---|---|---|
| Brown clay and stones | 4-6 | 4-6 |
| Gravel parting | 1-6 | 6-0 |
| Brown clay | 1-6 | 7-6 |
| Brown and grey shale, severely shattered | 6-6 | 14-0 |
| Grey shale, shattered | 47-0 | 61-0 |

C. 307.87 ft. (16359510)

| | | |
|---|---|---|
| Clay and stones | 3-6 | 3-6 |
| Dry sand clay | 3-6 | 7-0 |
| Clay and sand mixture | 7-0 | 14-0 |
| Very stony clay | 3-0 | 17-0 |
| Sand, gravel and clay | 8-0 | 25-0 |
| Clay, sand and gravel mixture | 11-0 | 36-0 |

D. 307.87 ft. (16359505)

| | | |
|---|---|---|
| Clay | 4-0 | 4-0 |
| Stony clay | 4-0 | 8-0 |
| Mixture of sand, gravel & clay | 11-0 | 19-0 |
| Sand and stones | 6-0 | 25-0 |

| | Thickness | Depth |
|---|---|---|
| E. 307.87 ft (16329500) | | |
| Clay | 3-6 | 3-6 |
| Clay and stones | 4-0 | 7-6 |
| Sand, gravel and clay mixture | 17-6 | 25-0 |
| PEN-Y-GELLI (SARN) (176905) | | |
| Brown soil | 4-0 | 4-0 |
| Gravel | 2-0 | 6-0 |
| Blue clay | 69-0 | 75-0 |
| Gravel | 29-0 | 104-0 |
| GARTHMML HALL (191989) | | |
| Earth and stones | 12-0 | 12-0 |
| Blue 'Pennant' stone | 4-0 | 16-0 |
| Clay and stone | 14-0 | 30-0 |
| Shale and stone | 66-6 | 96-6 |
| SHEET SO 29 | | |
| TIMBERTH (252966) | | |
| Clay | 8-0 | 8-0 |
| Hard shale | 102-0 | 110-0 |
| RHISTON (259951) | | |
| Soil | 2-0 | 2-0 |
| Stony blue marl | 18-0 | 20-0 |
| Soft blue shale | 8-0 | 28-0 |
| Stony blue marl | 20-0 | 48-0 |
| Blue marl | 56-0 | 104-0 |
| RHIW-GOCH (214966) | | |
| Soil | 5-0 | 5-0 |
| Shale and clay | 4-0 | 9-0 |
| Shale | 4-0 | 13-0 |
| MOAT (249969) | | |
| Clay and marl | 7-0 | 7-0 |
| Blue shale | 51-0 | 58-0 |

| | Thickness | Depth |
|---|---|---|
| **MIDDLE ALPORT** (275953) | | |
| Soil | 2-0 | 2-0 |
| Clay | 6-6 | 8-6 |
| Blue Shale | 91-6 | 100-0 |
| **SARN** (209909) | | |
| A. Soil | 1-0 | 1-0 |
| Hard brown stone | 10-0 | 11-0 |
| Big stones and clay | 11-0 | 22-0 |
| Grey shaley stone | 9-0 | 31-0 |
| Stones and grey clay | 12-0 | 43-0 |
| Grey stone | 3-0 | 46-0 |
| Clay and stone | 4-0 | 50-0 |
| B. Soil | 0-6 | 0-6 |
| Stone and rock | 4-0 | 4-6 |
| Broken rock | 3-0 | 7-6 |
| Gravel and boulders | 24-0 | 31-6 |
| Shale | 4-0 | 35-6 |
| Fine blue shale | 6-0 | 41-6 |
| Silty clay | 2-0 | 43-6 |
| Clay and shale | 14-6 | 58-0 |
| Hard blue shale | 16-0 | 74-0 |
| Blue shale and little clay | 19-0 | 93-0 |
| **CHURCH STOKE (THE VIEW)** (278939) | | |
| Stony marl | 12-0 | 12-0 |
| Hard blue shale | 52-0 | 64-0 |
| **CHIRBURY** (26059825) | | |
| Gravel and clay | 27-0 | 27-0 |
| Gravel | 11-0 | 38-0 |
| Blue clay and soft shale | 2-0 | 40-0 |
| **WOODLANDS, FORDEN** (221999) | | |
| Marl | 6-0 | 6-0 |
| Stony marl | 9-0 | 15-0 |
| Gravel and water | 6-6 | 21-6 |
| Marl | 5-6 | 27-0 |
| Shale | 8-0 | 35-0 |

| | Thickness | Depth |
|---|---|---|
| GAER BRIDGE (998214)[1] | | |
| Alluvium | 6-0 | 6-0 |
| Stony clay | 21-0 | 27-0 |
| Solid | | |
| SHEET SO 39 | | |
| MORE (343915) | | |
| Subsoil | 3-0 | 3-0 |
| Gravel, coarse, dry | 12-0 | 15-0 |
| Boulder clay with gravel | 20-0 | 35-0 |
| HARDWICK HALL (376905) | | |
| Stony marl | 40-0 | 40-0 |
| Gravel | 12-0 | 52-0 |
| Stony marl | 10-0 | 62-0 |
| Soft clay | 42-0 | 104-0 |
| Blue shale | 27-0 | 131-0 |
| CORNDON HILL (305975) | | |
| Clay | 12-0 | 12-0 |
| Gravel and grit | 6-0 | 18-0 |
| CORNDON HILL (305975) | | |
| Gravel, clay and shale | 8-0 | 8-0 |
| Gravel | 15-0 | 23-0 |
| Large gravel | 17-0 | 30-0 |
| Yellow clay and shale | 21-0 | 51-0 |
| NORBURY (365927) | | |
| Stony marl | 6-0 | 6-0 |
| Sand and gravel | 5-0 | 11-0 |
| Stony marl | 2-0 | 13-0 |
| Purple rock | 2-0 | 15-0 |
| THE HOME, BISHOP'S CASTLE | | |
| A. (377904) | | |
| Soil | 1-0 | 1-0 |
| Hard brown stony marl | 14-0 | 14-0 |

| | Thickness | Depth |
|---|---|---|
| Light brown stony marl | 10-0 | 25-0 |
| Soft blue clay | 2-0 | 27-0 |
| Hard light brown stony marl | 6-0 | 33-0 |
| Soft black mush | 2-0 | 35-0 |
| Soft brown stony marl | 7-0 | 42-0 |
| Soft grey marl | 8-0 | 5D-0 |
| Hard grey marl, soft layers | 10-0 | 60-0 |
| Hard brown shale | 4-0 | 64-0 |
| Soft grey clay | 55-0 | 119-0 |

B. (374904)

| | | |
|---|---|---|
| Soil | 2-0 | 2-0 |
| Brown stony clay | 2-0 | 4-0 |
| Soft light brown stony clay | 6-0 | 10-0 |
| Soft yellow,stony loam | 6-0 | 16-0 |
| Light brown clay | 4-0 | 20-0 |
| Soft light brown marl | 5-0 | 25-0 |
| Soft blue clay | 6-0 | 31-0 |

C. (373903)

| | | |
|---|---|---|
| Soil | 1-0 | 1-0 |
| Soft brown clay | 3-0 | 4-0 |
| Big gravel | 1-0 | 5-0 |
| Big stones with brown clay | 8-0 | 13-0 |
| Sandy loam/ water | 1-0 | 14-0 |
| Soft grey loam | 2-0 | 16-0 |
| Hard stony marl | 1-0 | 17-0 |
| Soft brown loam | 4-0 | 21-0 |
| Hard stony marl | 1-0 | 22-0 |
| Hard brown matted gravel | 2-0 | 24-0 |

D. (374901)

| | | |
|---|---|---|
| Soil | 1-0 | 1-0 |
| Brown clay | 3-0 | 4-0 |
| Big stones | 1-0 | 5-0 |
| Big matted gravel with clay | 7-6 | 12-6 |
| Big river bed gravel | 5-0 | 17-6 |
| Hard brown clay | 2-0 | 19-6 |
| Soft brown clay | 4-0 | 23-6 |
| Soft grey and brown clay | 5-6 | 29-0 |
| Soft grey clay | 20-0 | 49-0 |
| Hard grey stony marl | 11-0 | 60-0 |
| Hard grey stony marl | 26-0 | 86-0 |

| E. (373903) | Thickness | Depth |
|---|---|---|
| Soil | 1-0 | 1-0 |
| Brown clay | 1-0 | 2-0 |
| Soft yellow clay | 2-0 | 4-0 |
| Big stones and brown clay | 12-0 | 16-0 |
| Soft light brown clay | 3-6 | 19-6 |
| Gravel and loamy sand | 0-9 | 20-3 |
| Hard brown stony clay | 1-9 | 22-0 |

ROBURY RING, WENTNOR (39149302)

| | | |
|---|---|---|
| Pleistocene: Brown silty boulder clay | 14-9 | 14-9 |
| Silurian | 216-9 | 231-6 |
| Pre-Cambrian: W. Longmyndian | 21-6 | 253-0 |

SHEET SO 49

CHURCH STRETTON, HILLCREST (4649945)

| | | |
|---|---|---|
| Clay | 20-0 | 20-0 |
| Soft shale | 45-0 | 65-0 |
| Soft limestone | 7-0 | 72-0 |

CHURCH STRETTON, THE MALTINGS

| | | |
|---|---|---|
| Boulder clay | 89-0 | 89-0 |
| Hughley shale | 12-0 | 12-0 |

CHURCH STRETTON, SPRING BANK FARM (45719447)

| | | |
|---|---|---|
| Pleistocene: Head | 6-0 | 6-0 |
| Glacial sand and gravel, alternating with boulder clay | 80-0 | 86-0 |
| Silurian | 312-6 | 410-6 |

ALL STRETTON, INWOOD (46369674)

| | | |
|---|---|---|
| Soil | 1e0 | 1-0 |
| Matted gravel | 4-0 | 5-0 |
| Limestone (Longmyndian) | 175-0 | 180-0 |

ALL STRETTON, GORTSEYBANK (47L9671)

| | | |
|---|---|---|
| Subsoil | 11-0 | 11-0 |
| Dirty gravel | 30-0 | 41-0 |
| Sand | 30-0 | 71-0 |
| Marl (clay type) | 13-0 | 84-0 |

| | Thickness | Depth |
|---|---|---|
| Gravel | 3-0 | 87-0 |
| Marl (clay type) | 2-0 | 89-0 |
| Gravel | 1-0 | 90-0 |

SHOOTROUGH, NEAR COMELY (48899645)

(Inclined)

| | | |
|---|---|---|
| Pleistocene: Boulder clay | 26-6 | 26-6 |
| Upper Cambrian | 94-11 | 121-5 |
| Middle Cambrian | 118-1 | 239-6 |
| Lower Cambrian: Lower Comely sandstone | 565-8 | 805-2 |

BOTVYLE (47359653)

| | | |
|---|---|---|
| Boulder clays and glacial sand and gravel | 38-0 | 38-0 |
| Carboniferous | 60-0 | 98-0 |
| Silurian | 312-6 | 410-6 |

SMETHCOTT FARM (453996)

| | | |
|---|---|---|
| Stony marl | 37-0 | 37-0 |
| Sandy marl | 15-0 | 52-0 |
| Running sand | 1-0 | 53-0 |
| Marl with sand layers | 11-0 | 64-0 |
| Soft marl | 6-0 | 70-0 |
| Soft marl | 3-0 | 73-0 |
| Loamy clay | 13-0 | 86-0 |
| Brown shale and stones | 36-6 | 102-6 |
| Brown shale (loose) | 2-0 | 104-6 |
| Sand layers | 2-0 | 106-6 |
| Brown shale | 1-0 | 107-6 |
| Brown rock | 5-6 | 112-0 |

SMETHCOTT (457992)

(Taken across the river)

| | | |
|---|---|---|
| A. Blue clay and stones | 8-0 | 8-0 |
| Yellow clay and stones | 7-6 | 15-6 |
| Clay | 29-6 | 45-0 |
| Sand | 2-0 | 47-0 |
| Solid rock below | | |
| B. Soil | 1-9 | 1-9 |
| Gravel | 2-3 | 4-0 |

| | | Thickness | Depth |
|---|---|---|---|
| | Sand | 7-0 | 7-0 |
| | Clay | 23-0 | 30-0 |
| | Solid rock below | | |
| C. | Sandy Soil | 4-0 | 4-0 |
| | Loam | 2-6 | 6-6 |
| | Gravel | 12-0 | 18-6 |
| | Sandy clay | 7-6 | 26-0 |
| | Clay | 12-0 | 38-0 |
| | Sand | 3-0 | 41-0 |
| | Solid rock below | | |

SMETHCOTT (447989)

(Taken across river)

| | | | |
|---|---|---|---|
| A. | Clay and stones | 3-0 | 3-0 |
| | Gravel | 8-0 | 11-0 |
| | Clay and stones | 16-0 | 27-0 |
| | Gravel | 3-0 | 30-0 |
| | Hard sand | 14-0 | 44-0 |
| | Clay | 12-0 | 56-0 |
| | Blue clay and shale | 20-0 | 76-0 |
| B. | Clay | 3-0 | 3-0 |
| | Gravel | 4-0 | 7-0 |
| | Running sand | 14-0 | 21-0 |
| | Solid rock below | | |
| C. | Clay | 20-0 | 20-0 |
| | Loose rock, shale, clay and shale | 10-0 | 30-0 |

SHEET SO 59

CARDINGTON, GILBERRIES

| | | | |
|---|---|---|---|
| | Stony marl, boulder clay | 50-0 | 50-0 |
| | Pentamerus Beds | 25-0 | 75-0 |
| | Blue shale | 25-0 | 100-0 |

CHATWELL HALL

| | | | |
|---|---|---|---|
| | Marl | 4-0 | 4-0 |
| | Light green shale | 42-0 | 46-0 |
| | Blue shale | 7-0 | 53-0 |
| | Hard shale | 13-0 | 66-0 |

| | Thickness | Depth |
|---|---|---|
| Chatwell sandstone and flags | 59-0 | 59-0 |

EAST WALL FARM

| | | |
|---|---|---|
| Shale and marl | 25-0 | 25-0 |
| Blue shale | 30-0 | 55-0 |
| Grey shale | 35-0 | 100-0 |

SHEET SO 18

DOLFOR(105871)

| | | |
|---|---|---|
| A. Brown clay | 4-0 | 4-0 |
| Blue clay | 7-0 | 11-0 |
| Blue marl and stone | 7-0 | 18-0 |
| Blue shale and rock | 4-0 | 22-0 |
| B. Brown clay | 4-0 | 4-0 |
| Blue clay | 13-6 | 17-6 |
| Blue marl and stones | 16-0 | 33-6 |
| Blue shale rock | 3-6 | 37-0 |
| C. Brown clay | 4-0 | 4-0 |
| Blue clay | 4-6 | 8-6 |
| Gravel | 14-6 | 23-0 |
| Blue clay | 5-0 | 28-0 |

SHEET SO 38

ASTON-ON-CLUN (393817)

| | | |
|---|---|---|
| Clay | 29-0 | 29-0 |
| Clay and gravel | 5-0 | 34-0 |
| Clay (blue) | 46-0 | 80-0 |
| Shale | 5-0 | 85-0 |
| Rock, hard | 4-0 | 89-0 |

LYDBURY NORTH (354961)

| | | |
|---|---|---|
| Loam and gravel | 12-0 | 12-0 |
| Clay | 3-0 | 15-0 |
| Limestone | 45-0 | 60-0 |

COLEBATCH, THE PINES (322877)

| | | |
|---|---|---|
| Start 21-6 below ground | | |
| White marl | 3-6 | 3-6 |
| Yellow marl | 15-0 | 40-0 |
| Shale | 12-0 | 52-0 |

| BROCKTON (327858) | Thickness | Depth |
|---|---|---|
| Clay and shale | 7-0 | 7-0 |
| Gravel | 17-0 | 24-0 |
| Shale | 6-0 | 30-0 |

BROCKTON (alternative version)

| | | |
|---|---|---|
| Surface soil, clay and gravel | 4-0 | 4-0 |
| Gravel | 0-6 | 4-6 |
| Clay and gravel | 7-6 | 12-0 |
| Gravel and sand | 11-0 | 23-0 |
| Shale | 7-0 | 30-0 |

STAR FARM, COLEBATCH (320874)

| | | |
|---|---|---|
| Soil | 1-0 | 1-0 |
| Brown marl | 2-0 | 3-0 |
| Hard brown shale | 24-0 | 27-0 |

OAKLEY FARM, BISHOP'S CASTLE (33318811)

| | | |
|---|---|---|
| Gravel and loamy sand | 15-0 | 15-0 |
| Sandy clay | 3-0 | 18-0 |
| Gravel | 28-0 | 46-0 |
| Gravel and bands of boulder clay | 4-0 | 50-0 |
| Boulder clay | 1-0 | 51-0 |
| Blue clay | 13-6 | 64-6 |
| Boulder clay | 36-6 | 101-0 |

EATON FARM (375987)

| | | |
|---|---|---|
| Alluvium | 5-10 | 5-10 |
| Boulder clay with thin bands of gravel | 19-8 | 25-6 |
| Wenlock shale | 259-0 | 284-6 |

EATON FARM (37238978)

| | | |
|---|---|---|
| Pleistocene: brown silty boulder clay | 25-6 | 25-6 |
| Silurian: Wenlock shale | 232-6 | 258-0 |

SHEET SO 48

ACTON SCOTT (454899)

| | | |
|---|---|---|
| Clay | 30-0 | 30-0 |
| Shale | 10-0 | 40-0 |

| CRAVEN ARMS STATION(432830) | Thickness | Depth |
|---|---|---|
| Close gravel | 8-0 | 8-0 |
| Coarse gravel | 6-0 | 14-0 |
| Open gravel | 10-6 | 24-6 |
| **CRAVEN ARMS (43498325)** | | |
| Glacial gravel | 46-0 | 46-0 |
| **CRAVEN ARMS (434832)** | | |
| Sandy clay | 5-0 | 5-0 |
| Grey running sand | 6-6 | 11-6 |
| Sand and gravel | 1-6 | 13-0 |
| Grey sand | 6-0 | 19-0 |
| Sand and gravel | 1-0 | 20-0 |
| Sand | 4-6 | 24-6 |
| Dark loamy sand | 5-0 | 29-6 |
| Dark running sand | 15-6 | 45-0 |
| Blue clay | 3-0 | 48-0 |
| Blue shale | 1-6 | 49-6 |
| Grey shale | 4-6 | 54-0 |
| **LONG MEADOWEND, CRAVEN ARMS (410822)** | | |
| Boulder clay | 48-0 | 48-0 |
| Grey shale | 52-0 | 100-0 |
| Hard grey shale | 5-0 | 105-0 |
| **SHEET SJ 20** | | |
| **FORDEN, SPRINGFIELDS COURT HOUSE (243062)** | | |
| Clay | 40-0 | 40-0 |
| Hard shale | 52-0 | 92-0 |
| **FORDEN, CHURCH (227011)** | | |
| Stony blue marl | 24-0 | 24-0 |
| Blue shale | 76-0 | 100-0 |
| **LEIGHTON (243053)** | | |
| Soil | 1-6 | 1-6 |
| Marl | 18-0 | 19-6 |
| Blue shale | 70-0 | 89-6 |

| SHEET SJ 30 | Thickness | Depth |
|---|---|---|
| **MINSTERLEY CREAMERIES (376051)** | | |
| Made ground and alluvium | 19-0 | 19-0 |
| Boulder clay | 17-0 | 36-0 |
| Grey shale and limestone | 9-0 | 45-0 |
| **LOWER WALLOP (325076)** | | |
| Stony marl | 34-0 | 34-0 |
| Gravel | 22-0 | 56-0 |
| Soft marl | 34-0 | 90-0 |
| Yellow shale | 15-0 | 105-0 |
| **ASTERLEY (374071)** | | |
| Well | 25-0 | 25-0 |
| Blue shale | 18-0 | 43-0 |
| Fire clay | 8-0 | 51-0 |
| Blue shale | 23-0 | 74-0 |
| **SNAILBEACH (375022)** | | |
| Clay | 9-0 | 9-0 |
| Yellow shale | 38-0 | 47-0 |
| Blue shale | 54-0 | 101-0 |
| **WHITTON FARM, NEAR WESTBURY (344096)** | | |
| Stony brown clay | 9-0 | 9-0 |
| Soft blue clay | 41-0 | 50-0 |
| Shale rock | 15-0 | 65-0 |
| Very rough water-bearing gravel | 7-0 | 72-0 |
| Bright red hard marl | 8-0 | 80-0 |
| **MYTTONSBEACH (364004)** | | |
| Broken rock and clay | 5-0 | 5-0 |
| Light soft brown clay | 29-0 | 34-0 |
| Washed gravel and sand | 10-0 | 44-0 |
| Rock | 1-0 | 45-0 |
| **THE HILL FARM (308100)** | | |
| Soil | 3-0 | 3-0 |
| Clay | 3-0 | 6-0 |
| Shale | 5-0 | 11-0 |
| Rock | 58-0 | 69-0 |

| CAUSECASTLE FARM (339081) | Thickness | Depth |
|---|---|---|
| Yellow clay and shale | 15-0 | 15-0 |
| Blue shale (soft) | 90-0 | 105-0 |
| Very hard blue shale | 45-0 | 150-0 |
| Blue rock | 65-0 | 215-0 |
| **ASTON PIGGOTT (308034)** | | |
| Brown stony marl | 8-0 | 8-0 |
| **BINWESTON FARM (304038)**[1] | | |
| Late glacial gravels | 22-6 | 22-6 |
| Grey clay with many stones | 9-0 | 31-6 |
| Coarse sand grains mainly shale | 0-9 | 32-3 |
| Grey clay with many stones | 2-9 | 35-0 |
| Soft grey clay withnlayers of fine grit | 13-0 | 48-0 |
| Soft grey clay with silt and a few stones | 2-0 | 50-0 |
| Hard grey clay with abundant stones | 20-0 | 70-0 |
| Wenlock shale | | |
| **SHEET SJ 40** | | |
| **LONGDEN COMMON (438045)** | | |
| Clay and soil | 12e0 | 12-0 |
| Argillaceous shale | 16-0 | 28-0 |
| **DORRINGTON CREAMERY (479028)** | | |
| Subsoil | 3-0 | 3-0 |
| Dry gravel | 6-0 | 9-0 |
| Water-bearing gravel | 34-0 | 43-0 |
| **DORRINGTON (479031)** | | |
| Start 18-6 below surface | | |
| Clay with sand and gravel | 6-6 | 25-0 |
| Blue clay and small gravel | 23-0 | 48-0 |
| Hard grey rock | 3-0 | 51-0 |
| **DORRINGTON (476031)** | | |
| Sandy marl | 25-0 | 25-0 |
| Soft blue shale | 9-0 | 34-0 |

1. Rowlands, P.H. 1966, p. 13

| | Thickness | Depth |
|---|---|---|
| Clay | 26-0 | 60-0 |
| Blue shale | 18-0 | 78¾0 |
| Clay | 12-0 | 90-0 |

WALFORD HOUSE, DORRINGTON (476031)

| | | |
|---|---|---|
| Loamy sand | 35-0 | 35-0 |
| Coarse sand | 15-0 | 50-0 |
| Clay | 13-0 | 63-0 |
| Blue shale | 10-0 | 73-0 |
| Clay | 14-0 | 87-0 |

THE LEASOWES, WILDERLEY (443022)

| | | |
|---|---|---|
| Subsoil | 1-0 | 1-0 |
| Stony marl | 7-0 | 8-0 |
| Matted gravel | 9-0 | 17-0 |
| Sandy marl | 6-0 | 23-0 |
| Rock and marl | 111-0 | 134-0 |

NETLEY OLD HALL, DORRINGTON (466019)

| | | |
|---|---|---|
| Stony marl | 8-0 | 8-0 |
| Loamy sand | 1-0 | 9-0 |
| Stony marl | 4-0 | 13-0 |
| Matted gravel | 2-0 | 15-0 |
| Stony marl | 19-0 | 34-0 |
| Marl (soft) | 3-0 | 37-0 |
| Coal Measures | 1-0 | 38-0 |

SHEET SJ 31

MARCHE MANOR FARM, WESTBURY (335105)

| | | |
|---|---|---|
| Boulder clay | 20-0 | 20-0 |
| Sand | ? | ? |

CREW GREEN (328155)

| | | | |
|---|---|---|---|
| A. | Soil | 1-0 | 1-0 |
| | Stony brown clay | 3-0 | 4-0 |
| | Hard light brown marl | 3-0 | 7-0 |
| | Light brown shale | 10-0 | 17-0 |
| B. | Clay | 7-0 | 7-0 |
| | Shale and gravel | 7-0 | 14-0 |
| | Brown shale | 4-0 | 18-0 |

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